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MULTIPLE DIKES OF CAPE NEDDICK

BY JOHN C. HAFF

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ABSTRACT

Cape Neddick lies on the southwest coast of Maine in York County. The Kittery quartzite crops out around nearly the entire periphery of the cape and contains a large number of structurally complicated multiple dikes. The quartzite in this area has been intruded by a soda granite and by the Cape Neddick gabbro. Each of these intrusives contains its own suite of dikes whose members are not, however, involved in multiple relationship with one another.

Most of the multiple dikes strike nearly northeast, are nearly vertical, and are characterized by strong parallelism. Both structures and lithological associations were studied to determine features peculiar to these dikes. Certain types of behavior, for example, that called by the writer "braiding" and "intrasectate" and "extrasectate" combination, were found to be typical.

No structural peculiarities in the host rock which might have localized the multiple dikes could be detected. The condition necessary for their formation is believed to be the repetition of periods of essentially parallel fracturing in association with periods of magma availability. The study suggests that recurrent fracturing and contemporaneous intrusion of magmatic differentiates are probably strictly interdependent phases of the dike-building activity. The cause of this activity must eventually be sought in the behavior and influence of subjacent igneous masses previously active in the area.

INTRODUCTION AND ACKNOWLEDGMENTS

On Cape Neddick, on the extreme southern portion of the Maine coast (Fig. 1), dikes are extraordinarily numerous. Along the south shore of the cape they constitute more than 25 per cent of the rocks exposed. More than half of the dikes are multiple intrusions whose aggregate thickness is much greater than that of the solitary dikes. Because of their complex history, the dikes exhibit a great variety of structural features. Certain typical patterns of behavior, genetically associated with the multiple dikes, are very well represented and have been studied in detail.

Cape Neddick extends eastward for 1¼ miles from a point just south of the village of York Beach. Its outline is roughly that of an elongate oval with the long axis trending about 30 degrees south of east. Its protrusion at this place is probably due to the presence of a gabbro stock

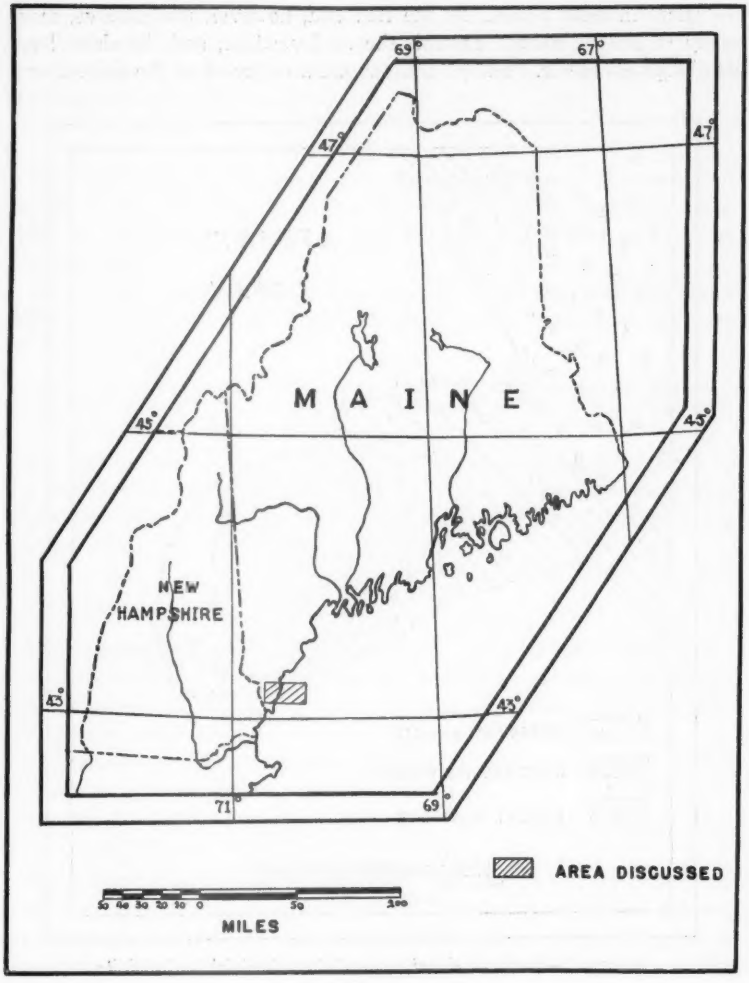

FIGURE 1.—*Index map*

that has invaded the quartzite country rock (Fig. 2). East of the cape, and separated from it by a narrow strait, is a small island called "The Nubble."

The greatest altitude along the backbone of the cape is less than 80 feet. The most prominent physical feature is the bluffed character of

the shore in most places. At the east end, however, the massive, more easily weathered gabbro controls the configuration, and the shore loses its abrupt character. The precipitous nature of most of the shore seems

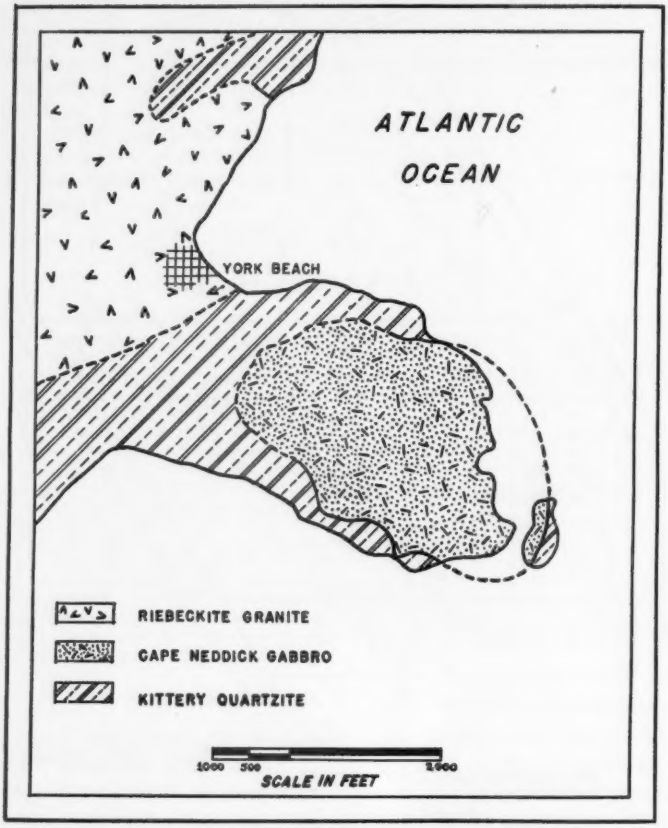

FIGURE 2.—*Geologic sketch map of Cape Neddick and vicinity*

to be due in part to violent wave attack, although the slabbing tendency of the steeply dipping quartzite seems also to have contributed.

The landward portion of the cape is almost completely covered with glacial drift, but toward the eastern extremity much bare rock is exposed, and outcrops are to be found on the shore at practically any tide.

The author wishes to acknowledge the assistance of Professor Charles

P. Berkey of the Department of Geology, Columbia University, whose observations in the field have proved very helpful. Both Professor Berkey and Professor S. J. Shand, also of Columbia University, have most kindly read and criticised the paper.

HISTORICAL BACKGROUND

The abundance and complex character of the dikes as shown by the exposures along the coast of southern Maine have been noted by Jackson (1838, p. 72), Hitchcock (1861, p. 214), Kemp (1890, p. 129), and Powers (1915, p. 172).

Katz (1918) presented a map showing the distribution of the major formations in the locality. With respect to the basic dikes encountered he states (p. 177):

> "Most if not all of them are younger than the post-Carboniferous granitic intrusives and are tentatively correlated with the Triassic eruptions of Nova Scotia and southwestern New England."

The granite and gabbro at Cape Neddick were assigned by Katz to the post-Carboniferous. Wandke (1922, p. 146) commented on some relatively rare paisanites, camptonites, and tinguaites associated with the alkaline complex in the region.

MAJOR FORMATIONS

QUARTZITE

The country rock is the Kittery quartzite. Katz (1918, p. 168) tentatively correlated the formation with the Merrimack quartzite of Massachusetts which is considered Carboniferous. Wandke (1922, p. 142) stressed the resemblance of the Kittery formation to the Isleboro formation of Rockland, Maine, which is believed to be Cambrian, but stated that:

> "The Kittery and Eliot formations are, however, provisionally dated as Carboniferous, following the lead of the U. S. Geological Survey."

The Kittery quartzite crops out almost continuously in a narrow belt along the shore of the cape. Because of the proximity of the gabbro, the dikes and quartzite are metamorphosed to varying degrees depending chiefly on distance from the contact; only toward the west where the gabbro contact curves northward are they relatively free from reaction effects. The strike of the quartzite varies from N 60° E to N 70° E except where locally disturbed. The dip is at places as low as 50 degrees but usually varies from 70 to 90 degrees.

The quartzite is typically thin bedded, and light and dark beds alternate frequently. Dark-purplish and chocolate-brown micaceous layers range in thickness from a quarter or half an inch up to 6 inches, and even

the thinnest laminae are easily followed. They show irregular pock marks and cavities. The more massive quartzose beds usually range from 1 to 5 feet in thickness. They are nearly white on the weathered surface, light blue and vitreous on the fresh fracture. They are uniformly fine grained, display fine banding, and many have an internal lenticular structure with individual lenses up to 10 feet in length. On the outcrop the quartzite is slabby and sheeted, owing in part at least to control by bedding.

The massive arenaceous beds show a well-defined jointing which is most conspicuous on nearly vertical faces at a small angle to the strike. The jointing is in three directions, and the quartzite breaks up into roughly triangular blocks. One broad crush zone involving large masses of fragmented quartzite was found on the Nubble at the contact of the gabbro and quartzite.

Folds of the quartzite ordinarily are represented only by the steeply dipping limbs of truncated forms. Close examination of certain beds shows tightly folded, highly contorted, isoclinal structures. The folding rarely involves the dikes themselves, and it may be inferred that the major folding antedated their emplacement. Abrupt swerves and shifting of dikes can almost always be referred to structures other than large-scale folding.

There are many indications of small but cumulatively important displacements. Small faults nearly perpendicular to the strike of the beds are common. Several periods of faulting are indicated by the number of such small faults and by the fact that the directions of apparent relative displacement are frequently reversed. The major fracture systems, such as those which sustain the pattern of the dike swarm and which control the directions of the transecting dikes, are quite divergent from the strikes of many of the smaller faults.

CAPE NEDDICK GABBRO

The Cape Neddick gabbro is a small stock-like body of oval cross section measuring approximately half a mile by three-quarters of a mile. The best exposures of its contact with the quartzite may be seen on the north side of the cape. On the south side the actual contact is usually concealed but closely parallels the shore road. The western boundary of the gabbro is concealed.

The dikes cutting the gabbro are usually measured in inches. Most of the smaller ones exhibit erratic, sometimes acutely angular deviations in course which are difficult to relate to any visible structure in the host rock. A post-gabbro dike of intermediate composition only rarely appears among the multiple dikes cutting the quartzite. Post-gabbro aplitic fluids have thoroughly permeated the surrounding quartzite, however, and appear as lenses, stringers, and as the matrix of one variety of breccia.

The members of the gabbro dike suite are in most instances solitary bodies. Occasionally more complicated relationships are seen in which two, or more rarely three, lithologies stand side by side. In such cases there are abrupt textural variations within the dikes, and age relationships are difficult to determine, for chill phenomena are absent. The classification of these "mixed" dikes is thus very uncertain; they cannot be classified as composite or as multiple dikes.

GRANITE

The typical riebeckite granite at York Beach is massive, uniformly coarse grained, and light buff. It is probably part of the Mount Agamenticus complex exposed several miles to the west. Only its northern contact with the quartzite is well exposed.

Four dikes cut the granite. A few feet from the northern contact is a fine grained, strongly porphyritic monchiquite 5 to 10 inches thick. About 300 feet farther south a narrow augite andesite, 6 to 8 inches thick, extends intermittently for more than 100 feet. Near the southern margin of the granite are two dikes of riebeckite microgranite about 4 to 5 feet thick.

MULTIPLE DIKES

DEFINITION

The terms composite dike and multiple dike, as used in this paper, have been defined by Bailey and Richey. A composite dike is characterized by basic members flanking a more acid central member. The basic portions chill against the country rock; the more acid central member shows no chill phenomena against the basic members. They remark (Bailey and Richey, 1924, p. 356):

> "In this memoir the distinction between multiple and composite intrusions is based upon the presence or absence of chilling at the contacts of the participants."

They accordingly define a multiple dike as a minor intrusion whose members, irrespective of composition, all chill against one another.

The term multiple intrusion signifies that the time interval between recurrent intrusions has been long enough to permit consolidation of each member before the next successive member is emplaced. As compared with composite intrusions multiple intrusions do not have symmetrical relationships between the components. Multiple dikes are further distinguished by the fact that nearly contemporaneous, chemically related magmas are not postulated.

DISTRIBUTION AND ATTITUDE

The most outstanding feature of the dikes in the Kittery quartzite is their predominantly multiple character. Not only are there many multiple dikes, but their components are usually numerous. One dike contains

16 parts with 11 lithologies. The more complex dikes also show a wide range of subordinate structures. The latter are of great assistance in developing a concept as to the unusual conditions responsible for these multiple dikes. One can observe a very detailed, though in places incomplete, series of the elsewhere less conspicuous physical features characterizing the dike-building processes. This is due in part to the excellent exposures, but more especially to an exceptionally complex history of deformation and intrusion. Furthermore, the most complex multiple dikes are usually found where the concentration of dikes is numerically greatest.

The narrowness of the shore and the smallness of the cape might possibly cast doubt upon the actual prevalence of such multiple dikes in surrounding areas. Many of the dikes in the quartzite strike nearly parallel to the bedding—N 60° – 70° E. Hence a traverse along the cape in a southeasterly direction crosscuts the strike at a high angle and affords exposures of the greatest number of dikes. On the northeastward-trending coast dikes are not encountered in such abundance since the direction of the coast is essentially parallel to the strike of the quartzite. Field evidence definitely shows, however, that multiple dikes are numerous in the surrounding regions and that their formation was not localized. However, special conditions, closely confined to the cape, have produced a greater concentration of dikes there. The nearly central location of the gabbro and the evidence of abundant, locally intense deformation probably account for the concentration on the cape. Moreover, the presence of breccia-bearing dikes, as well as significantly located large masses of breccia, as that on Nubble Island, indicates that the cape has probably been a focal point of repeated igneous activity of a much more intensive nature than has prevailed in the immediate surroundings.

So far as known no area containing multiple dikes in comparable abundance and complexity has ever been described. No other region along the adjacent coast line affords such a variety of structurally complicated dikes. The coast from Marblehead Neck, Massachusetts, to beyond Kennebunkport, Maine, has been thoroughly examined without finding comparable areas.

There are three distinct suites of dikes at Cape Neddick. Both the granite and the gabbro contain a single succession of small dikes. The multiple dikes in the quartzite represent several distinctly separated periods of activity. However, most of the dikes in the quartzite are older than either the granite or the gabbro. Dikes younger than either the granite or gabbro are relatively abundant in the quartzite only near the contacts with these bodies. The only exception to this is post-gabbro aplitic material which is represented by small dikes and stringers over the entire cape but seems to diminish in quantity toward the west. Only

one dike (a paisanitic type) known to belong to the post-granite series was found in the quartzite about a quarter of a mile from where the concealed granite-quartzite contact probably lies. Only the dikes in the quartzite will be considered in detail in this paper.

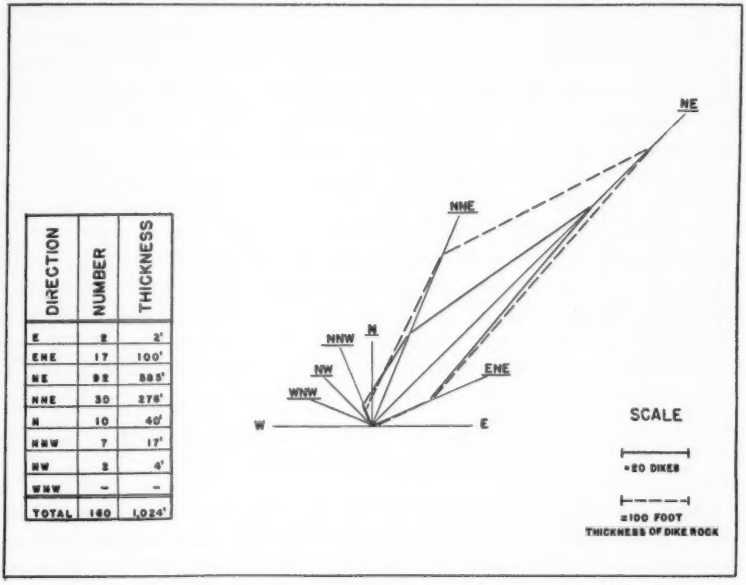

FIGURE 3.—*Diagram showing direction, number, and aggregate thickness of the multiple dikes on the south shore of Cape Neddick*

Note that maximum thickness is developed in direction of predominant strike.

Specimens of 338 different minor intrusions were collected within the quartzite on a traverse of the whole shore line. This figure is not the true total, for exposures along the shore are not complete. No great error can be involved in this figure, however, because the ratio of covered areas is relatively insignificant. Of the 338 intrusions no more than 5 per cent belong to the post-gabbro and post-granite suites. Some of the members of the swarm of 338 intrusions are solitary dikes, but the greater number form multiple dike complexes. On the south shore alone there are more than 240 intrusions. Of these 160 form multiple dikes; Figure 3 shows their strike frequency and total thickness. Their aggregate thickness is slightly more than 1000 feet. They are found along approximately

4000 feet of shore line. Hence in this distance the multiple dikes alone constitute at least 25 per cent of the rock exposed.

The dikes are all essentially vertical. Single dikes of substantial size with a small angle of dip are rarely seen, and the multiple dike units

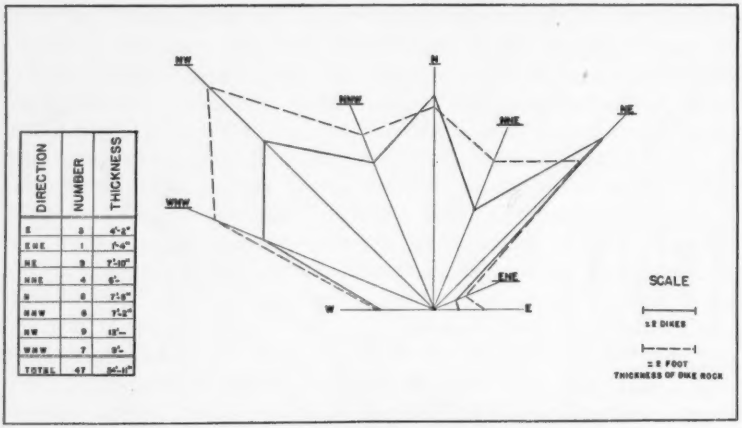

FIGURE 4.—*Diagram showing direction, number, and aggregate thickness of dikes cutting Cape Neddick gabbro*

hardly ever depart more than 10 degrees from vertical. Most of the dikes strike between N 40° E and N 45 ° E and make relatively small angles with the bedding of the quartzite. The small variation in strike between the dikes and the quartzite is not an indication of any tendency to a sill-like development, for the dikes cut directly across the folds of the host rock.

Such uniformity in disposition is not found in the dikes within the gabbro (Fig. 4). Forty-seven solitary dikes in this formation have a total thickness of about 55 feet. Although a few of these trend east and east-northeast, the others strike in various directions with about equal representation in the other sectors. Hence the pattern of the dikes in the gabbro is definitely radial. In the quartzite the greatest thickness of dikes has been developed parallel to their predominant strike direction. This is also true of the gabbro suite, but the correlation between predominant strike and aggregate thickness is not nearly so strong.

NOMENCLATURE

As used in this paper, the term suite is applied to a dike rock succession whose components are believed to represent a single epoch of

igneous activity. A complex is composed of a number of dikes representing different epochs of intrusion intimately involved so as to form a definite structural unit. The term component is used for one dike—one lithologic unit—in a complex. The term member is applied to one structural unit of any component of a complex. If, for example, a dike has been divided by the intrusion of a later one, either portion of the partitioned component dike may be called a member. The term epoch is applied to a period of dike-building activity within which a well-demarcated suite of dikes has been emplaced. Stage indicates a more or less well-defined period of time in the evolution of a complex. A given stage is usually detected by observation of transections, chill margins, and structural modifications which may demonstrate the aspect of a given complex at a particular time in its history.

Where whole suites or components of different suites are intermingled, the more inclusive term group is used. In this area two distinct suites, definitely related to the granite and gabbro respectively, can be differentiated. Some of the dikes and complexes in the quartzite belong to an early group characterized by multiple relationships. The members of this group are thought to represent more than one suite. A later, fairly well defined suite of dikes also appears in the quartzite, whose components habitually transect at large angles the dikes of the earlier group. In this region further accurate subdivision of the earlier group of dikes into sharply defined suites is neither feasible nor legitimate when the degree of their involvement and modification is considered.

During this study it became necessary, for descriptive purposes, to distinguish two types of multiple dikes. For clarity the discussion immediately following is restricted to the simple case in which only two dikes are combined to form a multiple one. In a multiple dike the younger component lies either within the older or has been intruded along the contact of the older dike and the host rock. The fundamental distinction between these two varieties depends solely on the location of the channel of the younger component. If the younger is intruded within and parallel to the older, their age relations may be determined, for the younger chills against the older rock which has been partitioned. Where the channel of the younger component has been along the contact between the older dike and the country rock, the chill margins of each component will be contiguous, and the succession cannot be determined.

The relationship of two dikes in a multiple complex is said to be intrasectate when the younger lies within the older and hence definitely shows their relative ages. The relationship of two dikes in a multiple complex is extrasectate when the later member lies outside of, but con-

tiguous to, an earlier dike, in which case they chill mutually against one another, and the succession cannot be determined.

The term intrasectate is restricted to the relationship between two dikes, but any number of contiguous dikes which chill against each other are extrasectate. It is not implied that the members of a multiple complex must have either an intrasectate or extrasectate relationship throughout the extent of their combination as a unit, and hence the basis of distinction is not really genetic. However, in determining age relationships and for descriptive purposes, this type of discrimination has proved indispensable.

A multiple dike consisting of many parts may present practically unlimited combinations of components showing various intrasectate and extrasectate relationships. In certain fortuitous cases a highly symmetrical development of intrasectate dikes is found, and their relative ages are perfectly obvious. At the other extreme dikes are frequently found in which all the members are extrasectate. One is therefore forced to use inference, based perhaps on a knowledge of the succession in adjacent complexes, in deciphering ages. Usually component dikes of a multiple unit are intergrown in various combinations of intrasectate and extrasectate character. Such complexes as the latter, with which are usually combined dikes transecting at large angles, must most often be depended upon in working out details of the succession.

MODES OF COMBINATION

The structurally and lithologically complex multiple dikes exhibit many distinctive combinations of their members. Different modes of combination are usually most clearly shown in dikes with relatively few members. As the multiple unit becomes increasingly complex, its structural history during earlier periods of development is apt to be obscured. The following examples of observed combinations are intended to indicate the general range. They also show that several different sets of conditions have prevailed during the development of the complexes, which are usually designated "multiple" without further division.

A common relationship is that in which the later dike or dike complex transects the earlier at a large angle (Fig. 5A). In a similar type of combination the earliest dike is cut by a second one which transects the older at a large angle; this simple group is then cut by another dike having an intrasectate relationship to the earliest one. At most, the truncated ends of the second dike are usually only very slightly offset (Fig. 5B). The offset, the absence of any recognizable reaction effects, and the presence of strong chill margins in the youngest dike with respect to the crosscutting one are noteworthy. These facts indicate that we are

here dealing not with any process of attack by incoming magma, but with one controlled wholly by structure.

Certain examples show a sinuous course of the intrasectate dike (Fig. 5C). In dikes with this type of combination search was made for physical features indigenous to the earlier dike which might have controlled the course of the opening followed by the later one. No structures were found, however, which could have been related to the distribution of the later dike. No joint systems in the earlier dike capable of controlling the later intrusion could be discovered. The thickness of the younger dike in this type of arrangement is usually quite constant. This suggests that the movements involved brought about a gaping or dilation of the flanks of the older dike. Had there been longitudinal movement of the two parts of the earlier dike, some parts of the adjacent portions would have been closer together than others, and a subsequently consolidated dike would have been of varying thickness.

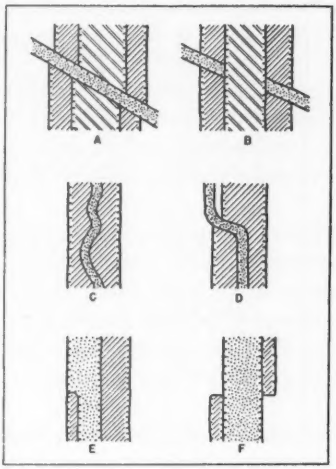

FIGURE 5.—*Typical modes of combination of multiple dikes*

A type of structure called in this paper "braiding", to be described more fully later, will be briefly noted here. Many dikes contain two components in which the intrasectate member (Fig. 5D) changes its course and cuts across either flank of the older dike. It thereafter emerges to become an individual dike, either permanently or usually for a relatively short distance. In most instances the younger dike does not continue for any great distance on a divergent course at a large angle to the older dike, but changes its direction so that it continues parallel to the older body or re-enters it. The repetition of such behavior gives to a large dike complex a very characteristic interwoven pattern. The preservation of the rectilinear character—*i.e.*, the lack of strong offset of the walls of the older dike on either side of the emergent one—shows that the walls have been only slightly displaced laterally. In any case the continuity of the opposite wall of the older dike indicates that no fault has passed directly through it.

In a type of structure transitional between an intrasectate and an extrasectate relationship, an intrasectate dike thickens at one place and

shoulders its way past the flank of the older body (Fig. 5E). The younger dike then becomes extrasectate. Although only a few examples of such structures were seen, no definite tendency of the younger dike to pass through the narrowest flank of the older body could be discovered.

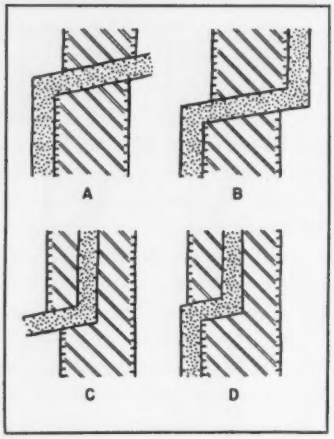

FIGURE 6.—*Examples of expectable changes in structural relationships of multiple dikes*

(A) The later contiguous (extrasectate) dike becomes transgressive to the older one, cuts directly through, and continues on the new course; (B) The elsewhere contiguous (extrasectate) dike becomes transgressive to the earlier, cuts directly through, again changes course and becomes extrasectate, but on the opposite side; (C) The transgressive dike changes its course while in the host, becomes intrasectate, and produces a multiple dike; (D) The extrasectate dike becomes transgressive, alters its course in the host, and becomes intrasectate. The multiple character is only temporarily lost.

In a common type of involvement (Fig. 5F) it appears as though the older dike has definitely been faulted across the strike, and that this episode was later followed by the intrusion of a second parallel dike. A displacement across the strike equal to the thickness of the older dike must necessarily be postulated to account for the observed relationship. Since most of the dikes thus displaced appear to have been thin, this may explain why it is so difficult to detect evidences of movement in the adjacent rock. It is also possible that displacement and emplacement were contemporaneous, and that the younger body has ruptured and forced apart the ends of the older dike. If the older dike were locally constricted and the zone of fracture developed along one wall, then crossed and permitted insinuation of magma between the opposite wall and the host rock, a somewhat similar relationship might result.

Some of the simpler combinations represent typical variations expectable in the multiple dikes and indicate especially well how changes in distribution and relationship take place. Some of these minor features are shown in Figure 6.

The operation of forces active during periods of tectonic activity depends largely upon the internal, relatively small-scale features of the rocks involved. This is of special significance when very small homo-

geneous areas or units are involved. Some of the responses observed in these complexes are very difficult to interpret. The first type of evidence that one seeks is indication of inherent weaknesses in the dikes involved, or the presence of structures developed subsequently to the consolidation, which might have controlled the mode of combination. In many instances none of these controls can be detected.

LITHOLOGICAL AND STRUCTURAL COMPLEXITY

Complexity in a multiple dike results either from an abundance of lithological varieties or because of structural involvement of its members. Striking examples of a successive compounding of lithologies as well as extreme structural complexity are found. The implications of these two sets of conditions are somewhat different.

The expectability of complex structural relationships increases toward the contact between the host rock and major intrusives. It also increases where there has been an overlapping of deformation effects in adjacent regions characterized by a different tectonic history. Further, the history of a given region may be an important factor contributing to complexity inasmuch as repeated deformation may develop a maze of accessible structures.

Lithological complexity depends more directly upon magma changes and variations and in a measure records the pulsation both in availability and composition. Multiple dikes showing a simple parallelism of components indicate only repeated fracturing in one direction. Development of zones of parallel or subparallel fracturing might be confined to one tectonic episode. The fractures developed in one episode may receive substances either from one magma or possibly from several, or the development of fracture zones may be intermittent, each stage being nearly contemporaneous with successive magma invasions. Although cooling effects indicate lapses of time between successive intrusions, such information is only qualitative. In some relatively simple complexes, however, inferential data may be derived from a study of the lithologic changes and the petrographic relationships of the rocks. The very simplicity of the structure in such a unit compels one to rely more strongly on petrographic evidence, inasmuch as such study may disclose facts concerning the evolution of the magma, assuming only one magma to have been involved. In more difficult cases an overlapping of effects of successive invasions, separated from one another by sufficient time to permit development of characteristic chemical and petrogenetic differences, is obtained. If such diagnostic features can be isolated, inferences regarding the influence of different magmas may be legitimately drawn.

The complex illustrated in Figure 7 shows exceptional complexity and

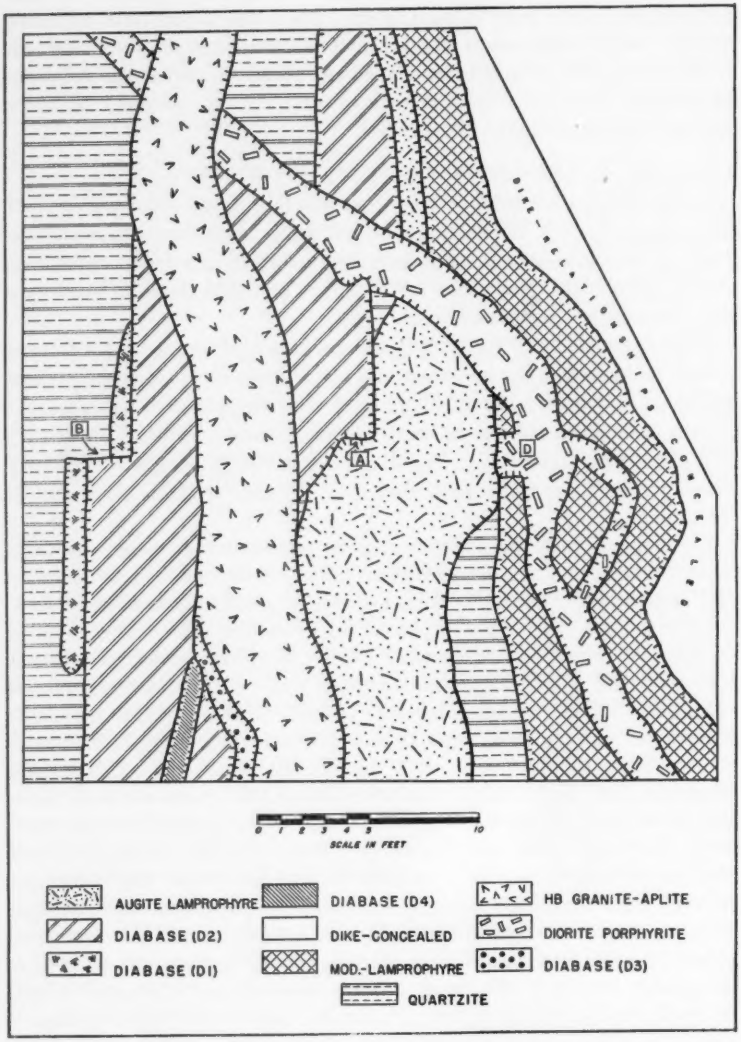

FIGURE 7.—*Complex showing effects of minor structures on various components*

exemplifies the effects of a series of movements on the form of successive dikes; that is, the present aspect has resulted in large part from the control of later dikes by structures existent in the earlier ones. The minor, very small-scale features in this section are especially revealing, and the process of complication can often be directly correlated with the structures of the earlier dikes, which in their turn responded to the structures of the host rock. The successive dikes are in addition unquestionably modified in form by structures developed contemporaneously with their intrusion. The channels of access for the different magmas have become more and more involved inasmuch as the later structures have also modified the earlier ones. The minor structures of the individual dikes of this complex indicate to some extent, although the record is incomplete, the sequence of movements in the evolution of the group; however, no significant indications of these are found in the quartzite.

The short, narrow diabase (D1) at the left of Figure 7 has been offset by faulting as is shown by the fact that the blunt termination at the re-entrant "B" is not chilled at its contact with the larger diabase (D2). The latter has not been faulted here, for that portion which abuts directly against the quartzite as well as that in contact with the truncated end of the thin diabase (D1) is distinctly chilled. Further, the blunt abutment of the diabase (D2) at "A" strongly suggests that the fault producing the offset at "B" may also have exerted influence on the outlines of the post-fault dike (D2) at "A". At the top of the figure diabase (D2) has been displaced to the right relative to the major part. The movement producing this offset may be related in time to the intrusion of the porphyrite, but it seems equally probable that this displacement occurred earlier inasmuch as several events intervened before the intrusion of the porphyrite. At point "D" in the porphyrite a projection transects the highly modified lamprophyre and terminates abruptly against the augite lamprophyre. Its location may be due to the fact that the porphyrite has taken advantage of the same plane of weakness along which the diabase (D2) was affected near "A". Thus this plane of weakness probably extended at least from "B" through "A" to "D" and possibly beyond.

The boundaries of the elongate block in the porphyrite are irregular. The porphyrite bifurcates and is then re-united so that it appears to engulf the block completely. A vertical section a few feet away shows that the porphyrite maintains its thickness with depth, and it may be that this mass is definitely columnar in shape and that it is completely disconnected for a considerable distance below the level of the exposure.

The short curved portion of the diabase (D3) lies in an expectable place and demonstrates a principle controlling emplacement which is fre-

quently overlooked. The contacts of the diabase (D2) and the granite aplite are curved and complementary. The formation of these multiple dikes is believed to be due to the repeated re-opening of fissures. The readjustments accompanying such re-opening will in all probability have components of movement along the strike and across the strike. When the course of a dike is curved and the re-opening takes place along its margin, a gap is created at this curve no matter what the direction of longitudinal movement. It is believed that the shape and location of the diabase (D3) resulted from adjustments along such a curved plane.

This unit is complicated both structurally and lithologically. It is with the aid of such exposures, however, that the regional sequence of intrusion can best be determined. The structures here are involved, and components frequently transect so that the history may be partially recapitulated both from direct observation and from inference. At least four distinct periods of intrusion can be deciphered, although the unit contains eight lithologies. Considered from the standpoint of lithologic development this complex shows a quite expectable trend from basic to acid magmas.

There have been at least two periods of faulting transverse to the major trend of the unit. This faulting has somewhat affected the form of the complex, but the dikes have nevertheless preferred the lengthwise fissures. It is difficult to evaluate the eventual effect this transverse faulting alone has had on the net configuration of the complex. The displacement at "B" is only a foot, but in this connection the availability of magma must always be considered. Although the regional strike of most of the large multiple dikes is definitely northeast, there are large dikes at variance with this trend. At least six periods of essentially parallel faulting, leaving the porphyrite out of consideration, have ensued, the majority being accompanied by considerable magma. This development of strong, frequently repeated, nearly parallel fracture zones and the accompanying flow of magmas account for the multiple character of the mass.

The assemblage in a given dike usually does not maintain, for any great distance along the strike, the lithologic distribution shown in a given outcrop. Figure 8 indicates some of the expectable changes in the distribution of members without increase in the number of components. In plan "A" a coarse diabase (D1) has been divided into three parts. At the right a thin augite andesite lies between the contact of the coarse diabase and the quartzite. About 30 feet from this section the complex is as shown in plan "B". The proportions in bulk of the different components are changed, and the relationship of the augite andesite is also changed. In plan "A" the diabase (D1) is 8 feet thick, whereas in plan

"B" it varies from 10½ to 12 feet. Although the behavior of the diabase (D2) is somewhat erratic, beyond the zone of its constriction it expands to a thickness of 5 feet, thereby retaining essentially its original proportions. The augite andesite, which shows no change in thickness, has been

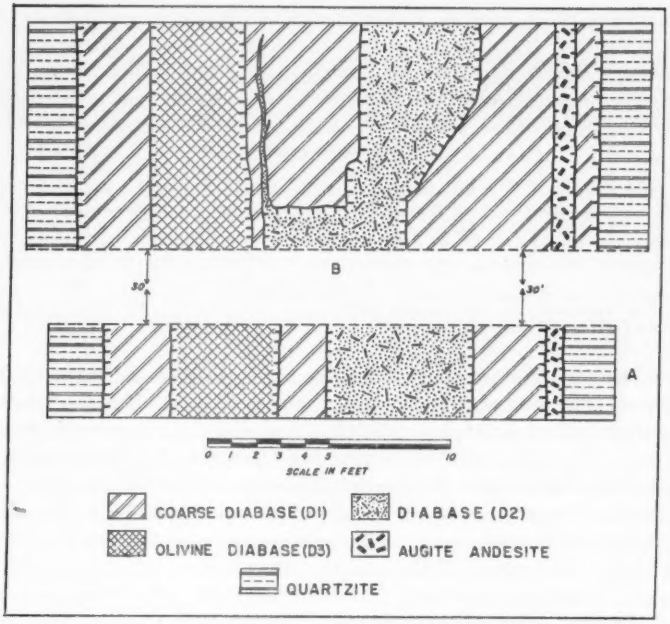

FIGURE 8.—*Complex of relatively simple structure*
Showing variations in lithologic distribution along the strike.

gently diverted so that in "B" it lies within the coarse diabase (D1). On the other hand, the outlines and general behavior of the diabase (D2) have been temporarily modified. Localized constriction has been followed by an expansion to nearly its former thickness within 5 to 6 feet. The distribution within the unit in plan "B" has also been changed somewhat, for within 30 feet the diabase (D2) has migrated until it lies within a few inches of diabase (D3). The offshoot from diabase (D2) has occupied an extension of the fracture which it followed earlier, indicating how one part of the channel has abruptly diminished in size and showing the earlier tendency to parallelism. As might be expected, no deformation can be detected in the host diabase in the disturbed area. There has been

no lateral displacement of the parts of the host dike along a fault plane, although the course of diabase (D2) was locally divergent.

These two plans show very well the changes to be anticipated along the strike in such a multiple dike. The component dikes everywhere exhibit minor variations in thickness and interrelationship. They wander

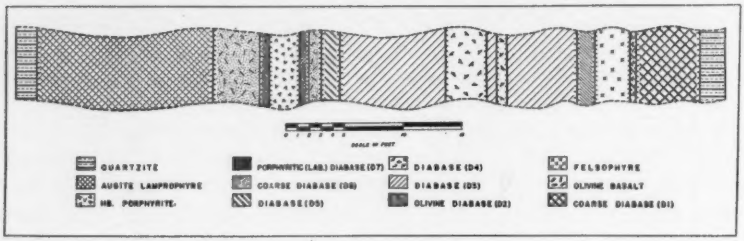

FIGURE 9.—*Multiple dike*

Showing lithologically the most complex assemblage at Cape Neddick. Within the limits of the exposure the structural relationships are unchanged.

back and forth through the walls of contiguous dikes and exchange host for host. The most significant feature, however, is that here the approach to parallelism of individual multiple dikes with respect to each other is not lost. Tapering, splitting, forking, and anastomosing behavior are entirely subordinate when the multiple dikes are considered as units.

The structurally simple but lithologically complex multiple dike shown in Figure 9 is the most complex assemblage observed. In the illustration the actual thicknesses of the smaller dikes had to be exaggerated, for some of them are only from 2 to 6 inches thick. This unit is not an isolated dike, for it is separated from neighboring dikes by only a few feet of quartzite. It is thus really an integral and highly representative part of the dike swarm but is unique in having such uniform structure and abrupt lithological alternations. The actual width of the dike is 55 feet; it is composed of 16 members and contains 11 different lithologies.

There are seven diabases, an olivine basalt, a very thick augite lamprophyre, a felsophyre, and a hornblende porphyrite. A much later felsophyre of very irregular course cuts this complex a short distance away without seriously disturbing the proportions here shown. The combined thickness of the diabases alone is 29 feet. Hence, slightly over 50 per cent of the mass is diabasic material, provided the relationships shown continue at depth. This unit substantiates findings concerning the general lithologic proportions in other complexes of the area. At other localities, however, many of the diabases are transecting in character. Although here the complex illustrates the usual lithologic proportions, no inferences as

to the age relationships among the diabases themselves can be made. The general feature of the abundance of basic dike material (not necessarily diabasic) throughout the region is also borne out, for only the small felsophyre is acid.

Eight dikes are extrasectate and three are intrasectate, indicating that here at least the zones of fracture tended to develop along the contact between the dikes and the adjacent rock. Although such a trend is well shown in this instance, it is not at present feasible to generalize concerning such preferential behavior throughout the region as a whole. Furthermore, the intrasectate dikes in general exhibit no tendency to bisect the host dike. The larger part of the diabase (D4) is nearly central in the diabase (D3). The coarse diabase (D6) has been invaded by another diabase (D7), which in turn has been intruded by the porphyrite. Diabase (D7) has cut the diabase (D6) so as to give the resulting doublet an asymmetrical character, but the later porphyrite has nearly bisected diabase (D7). From such relationships it is impossible to derive any satisfactory rule of behavior. Such relationships as are here shown may persist for 100 feet or more, perhaps only for 50 feet, but then almost certainly at least one component would become transecting, probably at a small angle to adjacent dikes. In such a case the distribution of the dikes within the unit would be somewhat changed, but the structural unity and lithological variety would not be destroyed.

Locally, the structure of this complex is so simple that despite its lithologic variety it gives few clues as to the relative ages of the components. The relative ages of the first five extrasectate dikes, beginning at the right in the figure, cannot be determined. It is clear that the porphyrite is younger than the diabase (D7), which is in turn younger than diabase (D6). The age relationships of any member of this triplet with respect to the other eight dikes, however, are unknown. Much of the record is thus missing. Only extraordinarily critical and diagnostic petrographic data would permit the determination of the relative ages of these dikes on the basis of microscopic and chemical criteria.

For purposes of discussion, structurally and lithologically complex multiple dikes have been considered separately. However, there can hardly be any doubt as to the interdependence of deformation effects and igneous activity and their frequent correlation in time. Changes in regional attitude, modifications of the original structures of the basement rock and of visible host rocks, and changes in geothermal conditions are known to be associated with significant changes in the equilibria and volume of subjacent magmas. At Cape Neddick the complex multiple dike units show these corollary effects, but as is expectable they are difficult to distinguish. Further, the exposures of a given unit represent a

relatively small area, and it is only when coordinated data from the entire region are available that valid generalizations can be made.

SHIFTING

In some exposures in the quartzite, dikes abruptly change strike or dip, or both, and sharply crosscut the bedding or other structures of the host rock. Such behavior, which is definitely not controlled by faulting, is called shifting. Sharp deviations are less common in dikes cutting the granite or the gabbro.

Such abrupt changes in dip (Fig. 10) illustrate the lack of geometrical regularity and the wavering ascending courses of many dikes. Most dikes in the quartzite have a very large angle of dip, but shifting, as seen in vertical sections, frequently results in an appreciable flattening at many points of their course, and, moreover, such shifting may be observed at several levels in the same dike.

Abrupt changes in inclination or strike which produce lateral shifting in the courses of the dikes usually result in the direct transection of quartzite bedding planes. The shifted dikes seen in this region show no tendency to follow the bedding planes of the quartzite near the loci of shifting. They are not diverted by folds, nor are there any phacoliths. It is difficult to understand why available weaknesses such as bedding planes were disregarded unless other planes afforded greater ease of penetration. The explanation of the observed facts requires a process which has destroyed the continuity of the quartzite and has, moreover, operated chiefly in a direction transverse rather than parallel to original structure.

The dip of the younger dike in Figure 10 has changed abruptly and with accompanying constriction at the same level at which the dip of the older dike changed. Apparently two fractures have intersected at obtuse angles, and the magma has followed them rather than the bedding of the quartzite. This is borne out by the fact that the older dike has been displaced slightly at the intersection, indicating definitely that there has been relative movement along the fracture occupied by the younger body. The movements causing the displacement of the older dike have effected a local flattening and constriction in the fracture occupied by the younger dike.

Shifting may take place with continuity as well as with real or apparent discontinuity. Both conditions are illustrated in Figure 11 in which the small dike shifts to the left at "A". In the lower part of the figure the same small dike is offset with apparent discontinuity.

Distinctive modifications of the form of the dike are usually found where shifting takes place. Constriction often occurs as is seen at "A"

in Figure 11 and also in the younger dike of Figure 10. As has been stated, such constrictions may frequently be due to displacements of the magma channel localized at the intersection of fractures. Moreover, the fracture zones occupied by dikes are normally somewhat undulatory.

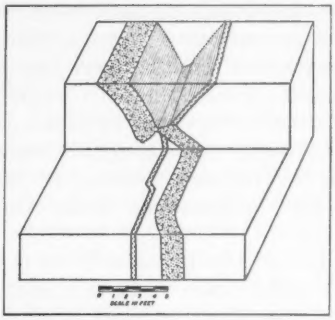

FIGURE 10.—*Block diagram showing shifting of dikes with abrupt changes in dip*

Note characteristic constriction of the younger dike at intersection.

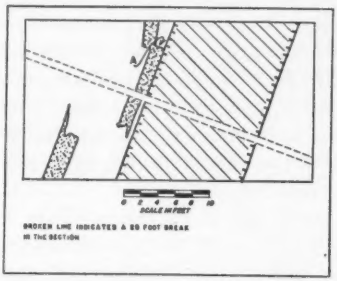

FIGURE 11.—*Plan view of dike shifting with continuity and apparent discontinuity*

Movements on their irregular surfaces, such as accompany intrusion, tend to modify locally the thickness and attitude of the dike. Different degrees of crushing strength, differences in the attitude of beds, and obscure weaknesses related to crystallization processes in massive plutonic rocks all influence the response of the rock to stresses. Where no visible, well-defined, large-scale structures are present, for example as in coarsely crystalline plutonic rocks, many inexplicable changes in the course of a dike must be related to internal weaknesses which cannot be detected by ordinary methods.

BRAIDING

Braiding is the interwoven structure of a dike complex produced by a predominant tendency toward parallelism of the individual dikes with respect to one another. Because the dikes of a complex are probably never parallel for long distances, but frequently are subparallel, they exhibit repeated transection and interweaving of the components. Braiding is especially typical of multiple dikes at Cape Neddick, where the fractures are numerous, closely spaced, and approximately parallel.

The size of the angle of transection varies, of course, and no limiting angle can be set forth, but an arbitrary figure of 20 degrees is assigned as typical for this area. Sharp bends are not usually found. The line of

approach of a braiding dike may be exactly along the contact of the dike transected. Rather than crossing directly many transecting dikes continue for some distance within the body of the older, in intrasectate relationship, before emerging. Such a dike may resume its former strike, or it may leave the transected dike and continue as a solitary one in a divergent direction. Two dikes may be parallel for some distance, whereupon one may swerve and either transect its neighbor or merely approach it, then turn away and resume its former strike. Such behavior may be extraordinarily variable within a relatively short distance. Typically the paths of the individual dikes of one of these interwoven complexes are tortuous, and component dikes may be braided with considerable variation in texture of the meshwork. When the braiding is of relatively fine texture, one may establish age relations easily, as small-angle transections are numerous.

Apparently the invasion of one braiding dike into another is not controlled by the primary internal structures of the dike transected. As has been mentioned, a braiding dike may enter another dike, continue within it for some distance in multiple relationship, then leave this dike and become a solitary dike again. Such behavior indicates that the course of the braiding dike is determined by a secondary rather than a primary structure of the invaded dike—*i.e.,* by the regional fracture system. The primary structural properties of the dikes themselves are not the cause either of the braiding or the tendency toward parallelism.

Braiding is rarely developed where large-angle transection occurs. It is expectable that separate braided dike complexes could form at large angles to one another, although only a few conspicuous examples were noted at Cape Neddick. Ordinarily the braided complexes are cut by non-multiple invading bodies.

Figure 12 represents one of the more extensive braided complexes on the north side of the cape. Diabases (D2) and (D1) show typical braiding. At one point diabase (D2) parallels (D1) and then, as shown in the lower part of the figure, enters the latter at a small angle. Without change of course it emerges, changes course slightly, and extends essentially parallel again. A very thin sliver of quartzite remains between the two dikes. Toward the upper part of the figure diabase (D2) sharply diverges from (D1).

A sudden shift in the course of diabase (D2), involving a constriction, can be seen at "X". The localization of this offset is probably related to the same minor structural control which produced the slight constriction in (D1). In contrast, where porphyrite (P1) shifts (near "Z"), there is no constriction. Such pronounced dislocation unaccompanied by constriction seems to be typical of offset along the more well-defined fault

planes. Where shifting takes place in response to a structure inconspicuous at present, there is usually noticeable constriction.

The diabase (D2) has been offset in three places. Each block has been moved in the same direction, but in varying amounts. Evidences of such small but probably cumulatively important movements are seen in many places. The attitudes of the separate portions of this diabase suggest the absence of any great thrusting accompanying the intrusion of the transecting dikes. Although it cannot be definitely determined whether or not the three faults were simultaneous, the fact that all visible portions are offset in the same relative direction seems to suggest that such was the case. On the other hand, such fractures as at "Z" and "Y" might not have been developed until the intrusion of the transecting dikes. Moreover, diabase (D1) has been displaced along the line of transection of porphyrite (P3), and diabase (D2) may well have been offset contemporaneously with the intrusion of the latter, as (P3) is later than either of the others. The disparity in the ages of the dikes cutting the diabase (D2) could thus be taken to indicate that it has been fragmented by separate movements related to different periods of intrusion. Although diabase (D2) is later than (D1), the displacements of the former are not all reflected in the latter. It is therefore possible that the effects of the displacement of diabase (D2) did not extend as far as the body of diabase (D1) (8 to 10 feet) or that the movements were somehow absorbed by it.

The porphyrite (P1) has been intruded along a fracture for some distance approximately parallel to the two earlier diabases. In this respect the general trend of the major elements of the complex is carried out even though, a few feet from the center of this ex-

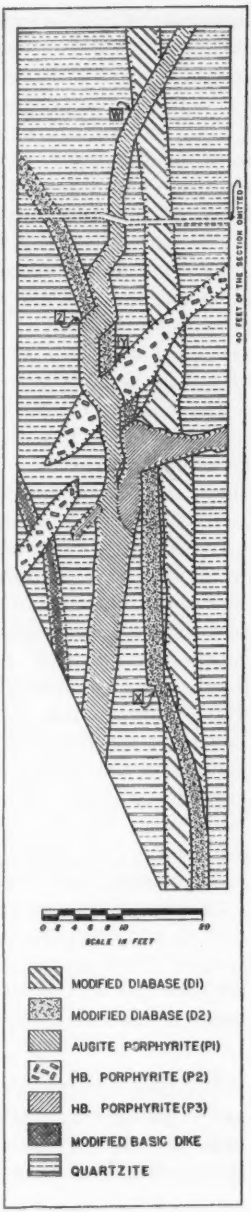

FIGURE 12.—*Braided multiple dike*

posure at "W", (P1) transects diabase (D1). Although the general parallelism of the major elements is conspicuous, there have been intervening periods of transection at a large angle to the stronger trend, inasmuch as the porphyrite (P2) is older than porphyrite (P1).

The two separate portions of porphyrite (P2) vary in thickness and taper conspicuously. Such irregular tapers, offset relative to one another, are common. In this case the two portions are about 6 feet apart. The larger portion expands considerably in crossing diabase (D1).

The age relations of porphyrites (P1) and (P3) could not be determined. As is usual near a concentration of dikes, and especially where similar lithologies are involved, confused contact relations may result. The two porphyrites weather so as to be megascopically identical. Frequently such complexes with irregular boundaries and general intermingling produce the false effect of a miniature stockwork of varying configuration.

The amount of involvement of the members of braided dike complexes at Cape Neddick is often greater than that shown. Where there are fewer intrusions, the pattern is an elongate meshwork of repeatedly interwoven dikes. A prominent parallelism in trend is maintained despite frequent intervening periods of transverse fracturing and intrusion.

In braiding the multiple dike components may migrate laterally in and out of the host dikes within a very short distance along the strike without departing from essential parallelism (Fig. 13). As seen in the vertical face there are two conspicuous irregularities in the outline of the fine-grained diabase (D2). One wall of this dike, at "Y", has contracted so that the dike is somewhat thinner there than below. At "X" the opposite wall expands. Diabase (D2) maintains its thickness after entering (D1), traversing the latter in intrasectate relationship. Diabase (D3) is, in vertical section, at first extrasectate with respect to (D2). At "X" this very thin diabase (D3) enters (D2), becoming intrasectate. Within a few feet the thin dike changes its course and becomes extrasectate with respect to (D2). Again changing course, diabase (D3) abandons (D2), at least locally, and becomes purely intrasectate to (D1).

Diabase (D1), one portion of which is host to the smaller dikes, is divided into two parts. Each portion tapers and is definitely arcuate in section. Their connection along the strike is not visible. The capricious behavior and the unpredictable, intermittent nature of the fracture zones, owing to control by local rock inhomogeneities, are well shown here.

STRUCTURES DUE TO FAULTING

Evidences of repeated faulting show that both solitary dikes and dike complexes have been involved. In the latter it is sometimes possible to determine the amount of horizontal displacement. It will be shown that faulting has, in at least one instance, obscured much of the evidence con-

cerning events in the development of the complex. The relationships illustrated indicate the danger of neglecting faulting when interpreting some of the lithologic alternations in the area.

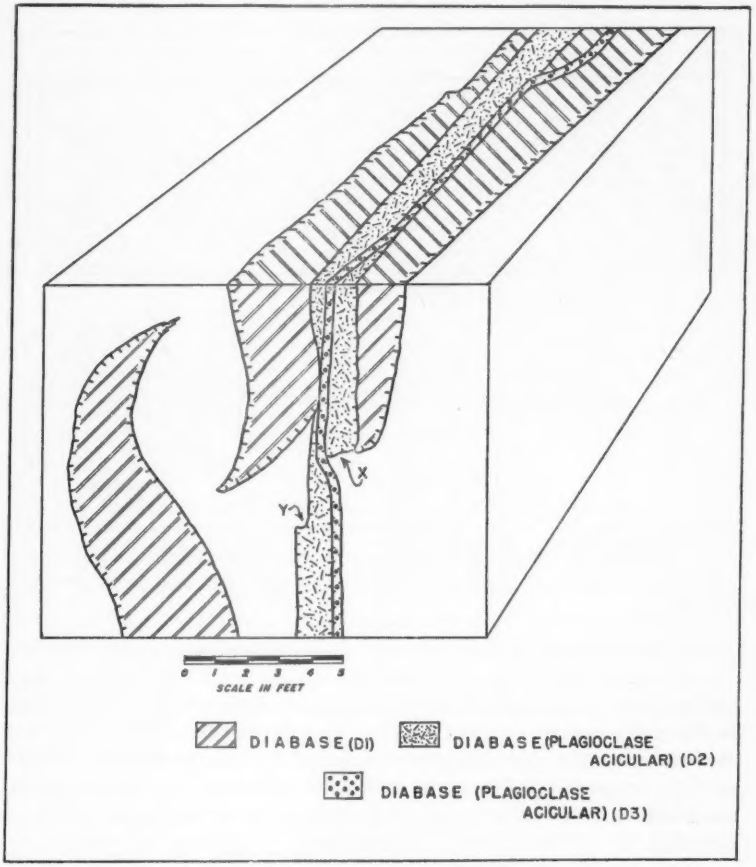

FIGURE 13.—*Braided multiple dike*
Showing the abrupt changes in relationship of the components.

Figure 14 illustrates a common manifestation of repeated faulting in which the relative displacements have been in opposite directions. The straight and strictly parallel sides of the large dike are unbroken. The small dike nearby, however, is offset 5 inches. The drag of this dike on

the fault plane may be clearly observed although no apparent displacement of the bedding planes of the quartzite is found. At "A", however, near the margin of the large dike, there is definite displacement of the quartzite laminae as shown in the inset. Although the movement at "A"

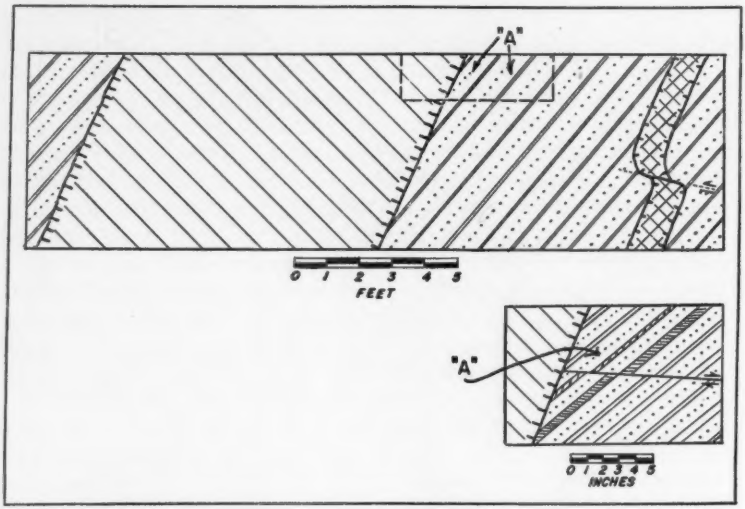

FIGURE 14.—*Faulted dike, accompanied by a large unfaulted one*

The straight margin of the large dike is taken as suggestive of a discrepancy in the ages of the two bodies.

is even less than that in the small dike, the relative offset is easily detected. The apparent displacement of these laminae has been in a direction opposite to that in the small dike. This suggests the age diversity of these faults and a disparity in the magnitude of the displacements. The fact that the large dike has straight walls whereas the smaller has been faulted strongly suggests a difference in their ages. Possibly the small dike was emplaced earlier, faulted, and at some later time the larger dike was intruded.

On the southwest shore of the cape a complicated exposure of highly modified dikes gives direct evidence of the magnitude of the displacements affecting the multiple complexes. Figure 15 shows a multiple dike, involving four lithologies, cut by a large, xenolith-bearing lamprophyre. Since the plane of the exposure here is horizontal, it is impossible to determine the amount of vertical displacement. However, the comparable thicknesses of the components of the xenolithic block and the unfractured

portions of the multiple dike suggest that the xenolith has not been moved any great distance vertically.

Where the lamprophyre (L1) crosscuts the multiple dike, their mutual boundaries tend to be notably irregular. Observation of the boundaries

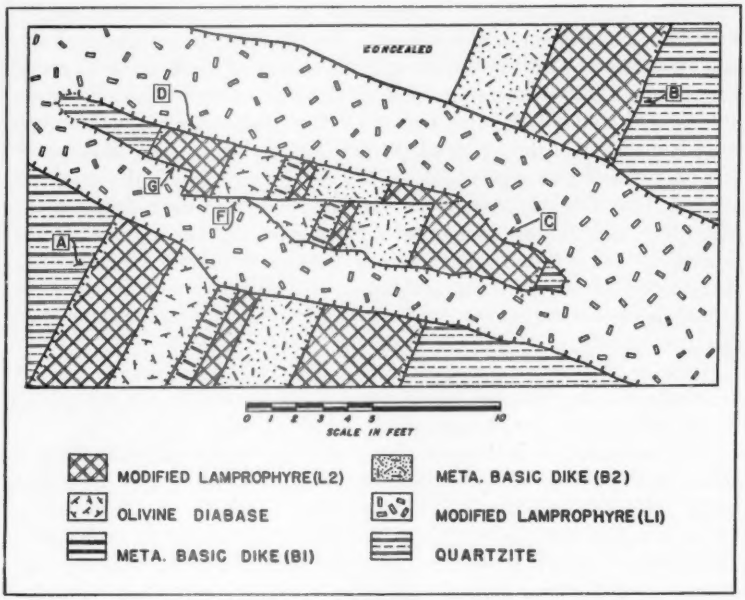

FIGURE 15.—*Large, xenolith-bearing lamprophyre cutting a faulted multiple dike*

of the xenolith, as compared with the transected ends of the blocks A and B, suggests that the margins of the xenolith have been modified. The boundaries of blocks A and B are no longer reflections of the outlines of the xenolith, although the latter, taken as a unit, has not been displaced laterally any great distance. Portion C of the faulted xenolith which has undergone greatest relative movement, as determined by its distinctly offset position with respect to block A, presents the more ragged boundary to the transecting lamprophyre.

The local crenulation of the contact of the crosscutting lamprophyre (L1) with the block C, if taken alone, might be interpreted as evidence of the emplacement of the lamprophyre by a process or processes combining assimilation and bodily incorporation of fragments accompanied by removal of portions of the older complex. On the other hand the

upper portion D of the xenolith has a straight boundary such as might be expected to result from faulting alone. The lower portion, however, has a crenulate and irregular contact with the transecting dike except where the inner fault plane forms the contact, as near "F". At "G" the form of the terminations of adjacent parts suggests that a block has been removed. The xenolith itself carries the most significant structure; it contains a fault which does not extend into the body of the lamprophyre, thus strengthening the probability of fault control of the boundaries under discussion. There is, therefore, structural evidence even within the xenolith itself that the earlier multiple dike has been faulted in a direction roughly parallel to the strike of the transecting body. The presence of this fault within the xenolith would not necessarily denote the existence of faults bounding it. However, evidence of such faulting is found in the fact that block B has been offset with respect to each portion of the xenolith as well as with respect to block A. The proximity of this multiple dike to the gabbro-quartzite contact (a critical area) and the presence of breccias in the region suggest that the movements here indicated were relatively strong and probably recurrent.

However, the faulted structure cannot be directly attributed to activity accompanying the intrusion of the gabbro. All these dikes are pre-gabbro in age and are cut by post-gabbro aplite stringers.

An illustration of structures that have been obscured by faulting is shown in Figure 16. Cross section "A" shows a multiple dike containing seven members and six lithologies. Five members are extrasectate, whereas a lamprophyre carries a hornblende porphyrite in intrasectate relationship. At first it would appear that this is a quite ordinary case of multiple intrusion.

Following along the strike of the complex from the section sketched in "A", it is seen that on one side of the hornblende porphyrite the lamprophyre wedges out. At this point a low irregular cliff is developed in the abrupt shore, and the relations in vertical section can be seen. The actual disposition of the components is very much different from section "A". At least three components, and possibly five, have been faulted. The movement has produced a relative lateral displacement of three members of about 8 feet, and apparently there has been only slight vertical movement. In the dismembering process the upper portion of the complex was moved laterally with only slight rotation, since the two opposed parts contiguous to the central porphyrite are still essentially parallel.

Below the level of the zone of dislocation (as at line of section B-B) the porphyrite is in simple extrasectate relationship. A horizontal section of the complex below the fault level would have been as seen in section

"B". Owing to the faulting, portions of two dikes have been obscured, as seen in horizontal section at the outcrop level, by the overlapping of the porphyrite.

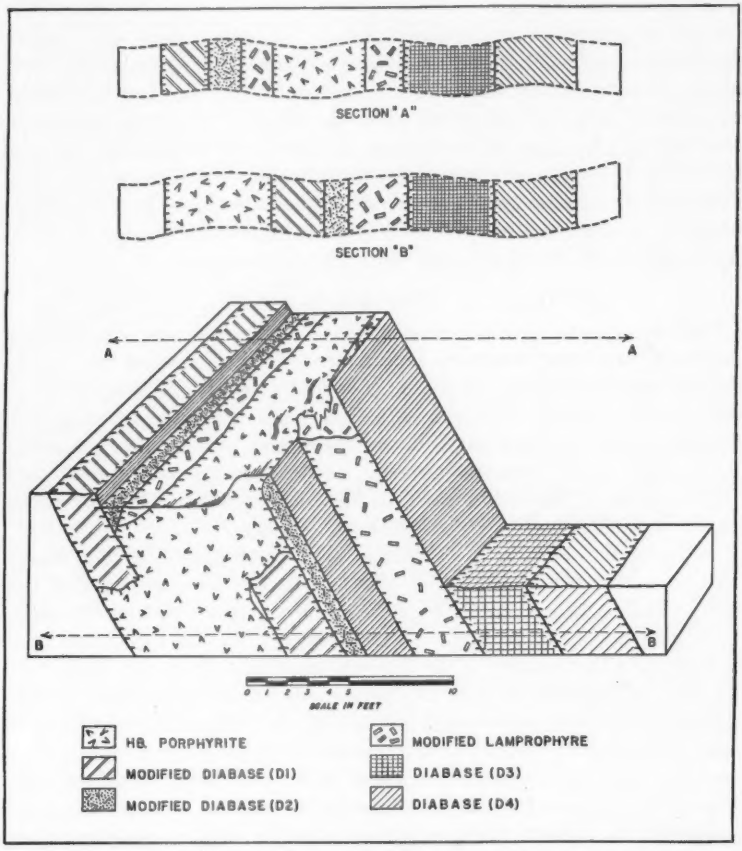

FIGURE 16.—*Multiple dike complicated by low-angle faulting followed by intrusion of porphyrite*

The outlines of the porphyrite in the face of the cliff suggest that its bounding walls were essentially straight before the fault zone was encountered. The fault cut sharply across the large diabase (D1), was then deflected along the wall of (D2), and cut the latter dike at a somewhat

higher level. No evidences of brecciation could be observed in the fault zone, nor are the sharply truncated ends of the displaced dikes conspicuously irregular in outline. This strike faulting did not take place until at least three periods of nearly vertical fracturing had allowed the combination of three dikes.

Such structures as are exemplified here indicate a much more complicated history than would have been suggested by cross section "A" alone. They definitely show that the development of multiple dikes may at times be involved with strike faulting, and that in certain cases such faulting may be required to explain the lithologic alternations seen in the field. It is of course recognized that multiple dikes are in the main developed by occupancy of nearly vertical passageways. It seems clear, however, that movements involving lateral displacement have, at times, intervened in the developmental history of the complexes.

TERMINATIONS

Various types of dike terminations have fairly characteristic outlines depending upon the structures encountered by the incoming magma in the different rock units. In the massive Cape Neddick gabbro dike terminations are usually well rounded and have a strong tendency to assume bulbous forms. In the quartzite, the pronounced structures of the rock, the attitude of the formation locally, and the substantial or relatively fragile nature of the individual beds have largely controlled the outline of the terminations, which are spiked and slivery.

Many of the dikes in the quartzite have extensions (Fig. 17) whose direction is often nearly parallel to the strike of the main mass of the dike. The prolongations in Figure 17 are of approximately the same thickness and nearly parallel. There is no evidence of the disturbance of bedding either outside of or between these prolongations. Such rectilinear development is taken as strong evidence of control by deformation structures and is especially convincing since the course of the main dike disregards the bedding of the quartzite. In the quartzite, dikes frequently abut squarely against massive beds. One might think that the massive character of the bed embracing the end of the dike acted as a buttress to prevent further extension, but in such a case local rock homogeneity is not always the determining factor since equally abrupt terminations sometimes cut directly across several lithologically distinct beds (Fig. 18). Where this occurs, no pronounced insinuation of magma along the readily available bedding planes was observed.

The terminations of the dikes in the gabbro show no strong tendency toward development of angular outlines but are either irregularly rounded or strongly blunted. Commonly the terminations may be slightly forked,

or the dikes may wedge out. Differences in the composition of the dikes do not seem to be a controlling influence affecting the mode of termination.

In abrupt, angular, and crosscutting terminations, the absence of brecciation is striking, and no dikes terminating in breccia were observed.

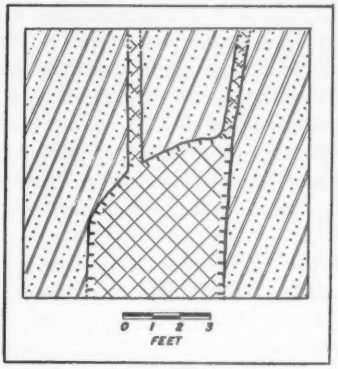

FIGURE 17.—*Blunt dike termination with narrow extensions cutting directly across bedding of quartzite*

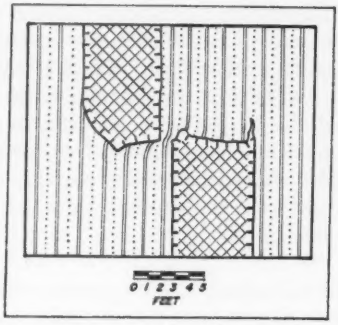

FIGURE 18.—*Offset dike*

The two portions terminate abruptly with only very slight insinuation of magma parallel to the bedding.

This suggests that the forces operative at the termination of the fracture were not adequate to shatter the host rock. Such a conception seems to be required, so far as observed, though shattering at terminations has been described elsewhere. However, branching and divergent terminations were found in the dikes in the quartzite. In such cases it seems probable that many small, irregular, and somewhat radial fractures were developed at the terminus of the fracture. Such a divergent fracture pattern probably had its inception in the stresses operative during the development of the master fracture, although there is also evidence that this condition was, in places, brought about more directly by the changing thermal conditions at the time of intrusion.

AGE RELATIONSHIPS

Detailed determination of the relative ages of many of the dikes and dike complexes in the quartzite is difficult, chiefly because of the usual parallelism. Further, although an intrasectate dike exhibits one kind of transection, the accompanying dikes of a multiple unit may be extrasectate and afford no information concerning their relative ages. All parallel dikes need not belong to the same period of intrusion, for there have been several periods of parallel fracture development and intrusion within the time range of the earlier dike group at Cape Neddick. Because of

parallelism, many dikes in a complex must be assigned to arbitrary periods, since one is unable to differentiate subordinate phases in the history of the complex.

In general and irrespective of where the dikes occur, age determinations may be made upon several different bases. Those conditions enabling one to discriminate between individual dikes will be discussed first. Ordinarily transection alone is employed, for, in the general case, if dike A is cut by dike B and dike B is cut by dike C, then dike C is the youngest. Such a criterion is unquestionably most satisfactory wherever available, but this is definitely not the ruling condition at Cape Neddick. In one prevalent kind of transection, however, the younger dike lies within the older, thereby presenting chill margins against the older body. This type of evidence is also unequivocal and is utilized in the case of intrasectate multiple dikes.

Chill margins are nearly always useful, except in the case of multiple dike members in extrasectate relationship. Chill margins are corroborative evidence in cases of direct transection and, in dikes of intrasectate relationship, indispensable to age determinations.

In the absence of transection, faulting alone may sometimes be of value in determining relative ages. If one of two, single, nearby dikes is faulted, a difference in age is indicated provided the dislocation extends across the path (not the body) of both dikes. Usually faulting can be detected in the rock adjacent to the intact dike. Thus since the course of the unfaulted dike totally disregards this fault zone, one may assume that the faulted dike is the earlier, even though their respective courses are essentially parallel.

The disposition of dikes with respect to folds and other structures in the host rock is of assistance in other districts, but this criterion cannot be used in the present study, for all the minor intrusives so far discovered are later than the regional folding.

In rare instances in this area the presence of breccias in a dike may give indirect evidence of the age of the breccia-bearing body with respect to dikes in the vicinity, for the breccia-bearing dikes may contain fragments of dike rocks identical with, or lithologically similar to, dikes in the region.

Inferences as to the relative ages of dikes may sometimes be drawn by comparing the degree of modification which different ones have undergone. When dikes are modified by shearing action, fracturing, and faulting, different degrees of change in mineralogy and internal structure result. However, criteria based upon apparently contrasting alteration effects must not be used without strong supporting evidence. Even then, a knowledge of the regional behavior and the typical responses of different

types of dikes must be known in some detail. Contact effects may likewise be severe, and these alterations may tend either to accent or obscure original differences.

To derive criteria regarding the ages of dikes applicable over this whole area requires the assimilation of all facts concerning the regional history, the behavior and structures of the minor intrusions, and the relationships of larger intrusives and the dikes associated with them. Only information so derived has proven to be of general use in determining the relative ages of dike groups and suites of Cape Neddick. The adjacent granite and gabbro each contain a suite of dikes. Since both these intrusives have invaded the quartzite, the later age of the post-granite and post-gabbro dikes with respect to the multiple dikes in the quartzite is conceded. The isolation of these dike suites permits one to establish criteria for their recognition and to determine, on a lithologic basis alone, the age of a representative when found in the quartzite.

The establishment of petrographic criteria for determining age relationships among the members of a multiple dike complex when all the member dikes are indigenous to one formation is more difficult. It may however be possible to distinguish petrographically suites of dike rocks if these suites were formed from magmas of widely different composition.

The strongest type of evidence indicating age differences to be found at Cape Neddick is a conspicuous discordance in both strike and behavior of the components of these complexes. The earlier group of dikes is characterized by a dominantly multiple habit and a constant strike. Superposed on the complexes of this group is a suite of transecting dikes, the components of which strike at a large angle to the earlier multiple ones. Further, these later dikes do not commonly enter into multiple relationships with one another. Hence there is a contrast between the two suites both in behavior and disposition. There is also evidence to show that, in the later suite, acid dike rocks are even rarer than in the earlier group. The discordance in strike, the submergence of the multiple dike-building trait, and the conspicuously different range of lithologies definitely indicate at least two distinct series of rocks. The dikes of the later suite in the quartzite very possibly have been derived and emplaced in one epoch of intrusion. However, the relationships between the components of the so-called earlier groups indicate possibly several periods of intrusion. The disruption of their structures by the later dikes and the commonly extrasectate character of the components of this early group make impossible an accurate discernment of sharply bounded, subordinate epochs of intrusion within the time range of this early group.

It is clearly established that there were several relatively early stages of basic intrusion now represented by the highly involved complexes

consisting entirely of basic dikes. It is impossible to make an accurate determination of the succession of the early basic dikes on structural data alone. On the basis of petrographic examination it seems probable that there were considerable differences in chemical character of the magmas producing the dikes of the earlier basic group. However, it is felt that satisfactory hypotheses regarding the evolution of the basic dikes can be obtained only with the use of chemical data. If such changes could be expressed quantitatively with the aid of chemical study, the course of differentiation between periods of intrusion might be determined in those complexes giving clear structural evidence of the succession. To do this, chemical criteria for each period would have to be established, since petrographically the composition of the dikes belonging to successive intrusions or members of a given complex does not appear to vary greatly.

Multiple dikes consisting only of basic members predominate at Cape Neddick. Acid dikes of substantial size are few. Complexes composed wholly of granophyres, quartz porphyries, and bostonitic rocks are not found. The multiple dike shown in Figure 19, in addition to its typical lithological associations, is notable for the clarity with which it shows many features of its evolution.

There are seven components in this complex, six diabases and one very small augite andesite dike. The regional history cannot be entirely reconstructed, for probably all the dike-rock lithologies intruded in the area in the time range of the complex are not represented. The age relationships of all of the members cannot be determined because of the parallelism of the earlier members, but sufficient information is available from this section to permit definite establishment of at least four stages in its development. Two of these stages are considered to be typical for the region as a whole.

(1) The relationships of the diabases (D1, D2, and D3) are believed to represent a very early stage. Since these three dikes are nearly consistently parallel, they may have been controlled by fracturing developed at one given stage.

(2) The intrusion of diabase (D4) followed the emplacement of diabase (D3). Assuming that the three dikes (D1, D2, and D3) belong to the same group, the small transecting dike (D4) is taken to represent a definitely later, second period.

(3) During this stage the complex was invaded by the porphyritic diabase (D5). This dike transected both diabases (D3 and D4). It then continued for a short distance in the quartzite and became extrasectate, lying between the margin of diabase (D2) and the quartzite. It is possible that diabase (D5) is older than (D2), but the transecting rela-

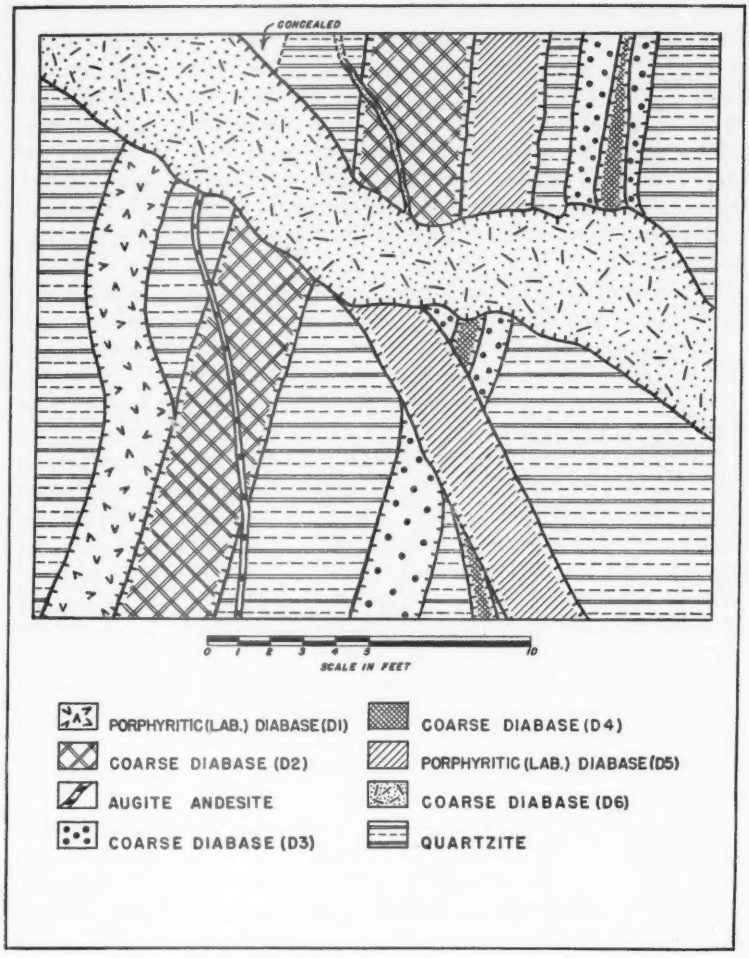

FIGURE 19.—*Multiple dike consisting entirely of basic components*
Showing four stages in development.

tionship of (D5) and its change in course where it abuts against (D2) before becoming extrasectate is taken as evidence of a hiatus of sufficient duration to usher in a new period of fracturing. It is also significant that diabase (D5) has transected diabases (D3 and D4) by making its entry along the same weakness zone utilized by diabase (D4).

(4) After the complex had reached this stage, the components were faulted with a displacement of 4 feet. After the faulting a new series of controls were operative, in accordance with which the large transecting diabase (D6) was emplaced. This youngest dike occupied a zone of weakness which extended through the area in which the earlier diabase (D5) changed its course from a crosscutting body to an extrasectate one. It is quite possible that the weaknesses, which earlier manifested themselves by inducing this change in the course of dike (D5), have, to some extent, persisted so as to influence the direction of a much later fault line.

The complex illustrated in Figure 20 contains a bostonite and a granophyre and is typical of the lithologically well diversified dikes.

The granophyre is extrasectate with respect to the adjacent dikes. However, elsewhere in the vicinity, granophyres are, like the bostonite, generally crosscutting. Probably neither the granophyres nor the bostonites appeared at all, or only rarely, in the earliest suites of dikes.

Morphologically the complex possesses some very interesting structures aside from lithology. All that can be definitely said concerning the age relationships of the acid members is that the bostonite and granophyre are younger than the large host diabase. It is unfortunate that the contact of the bostonite with the granophyre is concealed. At no other place in the entire area were these two rock types found in contact. However, the position and continuity of the slender tip of the porphyritic diabase (D1) is suggestive of its age with respect to the bostonite. It seems probable that if the porphyritic diabase (D1) were earlier than the bostonite, movements accompanying the emplacement of the latter would have destroyed the fragile extension. There is therefore some basis for supposing that the porphyritic diabase (D1) is later than the bostonite.

The xenolith of coarse diabase (D2) in the granophyre is of more than passing interest. Its presence is additional evidence that fragmentation is common where diabases and granophyres occur together in this area. Although such a condition can hardly be described as brecciation, we have at least another symptom of this tendency which is elsewhere very strong indeed.

The blunted portion of the porphyritic diabase (D1) represents nearly the same condition as seen elsewhere (Fig. 8). It emphasizes the relative frequency of such abrupt changes in thickness. As is common among these dikes, there is no corresponding reflection or offset in the margins of the host dike. This is one instance where a change in thickness has occurred without accompanying displacement or visible crushing along a fault. Accommodation of the porphyritic diabase (D1) appears to

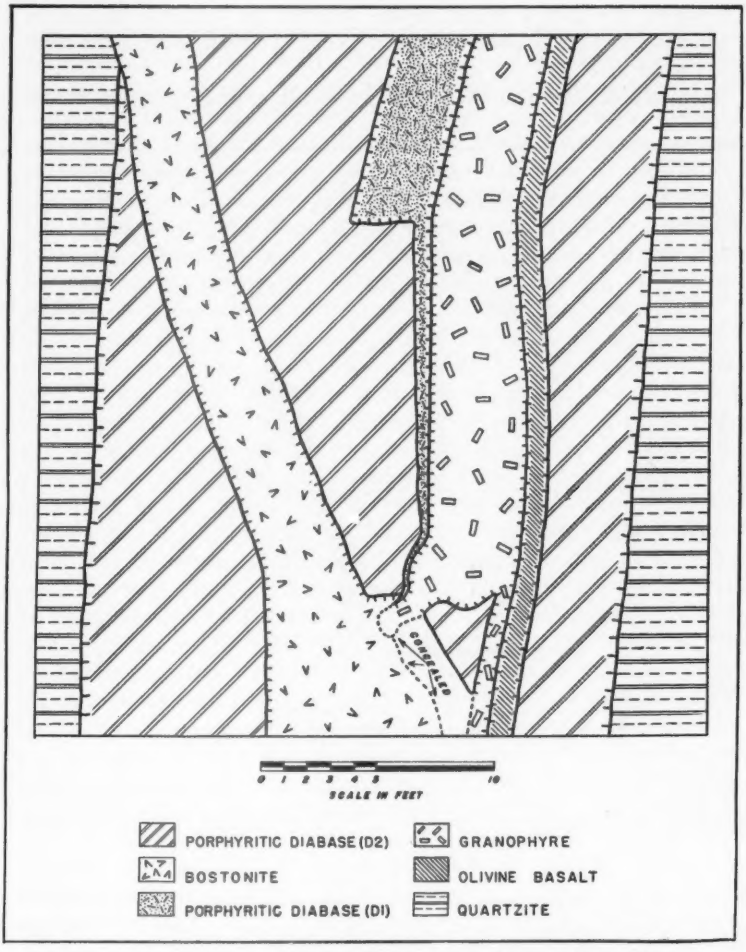

FIGURE 20.—*Multiple dike containing both granophyre and bostonite*
This is the only exposure in which these two rock types were found together.

have been made possible by the development, ultimately, of a clean fracture accompanied by a gaping, or expansion, of (D2) to make way for the incoming magma. As is indicated by the configuration of the two portions of the older diabase, some parts of it have been removed. The only real difference in the relationship here between the two diabases and

that shown in Figure 8 is that in this instance the narrow continuation of the younger dike is located along the contact rather than in the body of the older dike.

The complex shown in Figure 21 represents another example of the

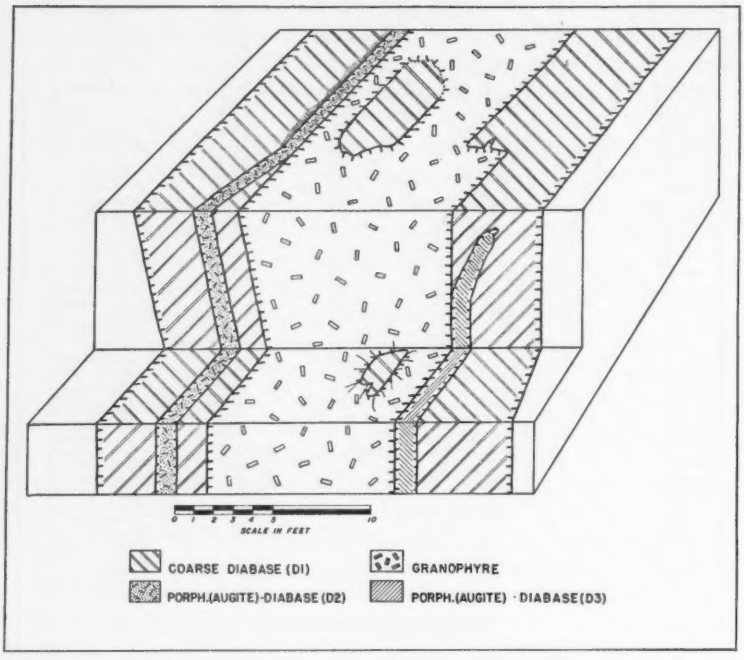

FIGURE 21.—*Multiple dike with large granophyre dike containing xenoliths of earlier host diabase*

association of diabase and granophyre; in this case, however, there is more definite fragmentation of the early basic dike and some hybridization by the acid magma. Hybridization is represented by a narrow, light-gray, mixed, albitized zone, rarely exceeding 2 inches in width, at the contacts between the larger diabase and the granophyre. Neither the walls of the large diabase nor the margins of the xenoliths have been much embayed.

Both small diabase dikes are considered to be later than the granophyre. In some parts of their courses both are extrasectate to the granophyre, so that chill effects alone do not suffice to establish their

age. Diabase (D2) exhibits normal chilling and shows no contact effects which can be ascribed to hybridization. Assuming that hybridization has not been selective, the mere absence of such phenomena along the margins of this small dike would indicate that it was younger than the granophyre. Since the small diabase has presumably been interposed between the older diabase and the granophyre, the former contact between the older bodies should show mixing effects. However, the areas in question (shaded in figure) show no indications of hybridization, elsewhere conspicuous at the contacts between the large diabase and the granophyre. It is quite probable that the channel of the small diabase resulted from movements in this zone which destroyed the old hybridized margin.

The same absence of mixing effects also characterizes the contacts of diabase (D3) and the granophyre, but there is additional evidence for inferring the age of the former. A xenolith of coarse diabase lies not far from the contact in question. If diabase (D3) antedated the granophyre, one would expect the latter to contain a portion of this small dike rather than of the larger diabase, granted that this xenolith came from the immediate vicinity. If (D3) is pre-granophyre in age, one must postulate the removal of the xenolith from behind the wall of porphyritic diabase. This removal would have to have been accomplished at another level since the form of the small diabase is here unmodified. Such a process is not impossible if one keeps in mind the frequent changes in thickness and shape of some of these bodies and especially the tenuousness of some of the thinner dikes. In the absence of definite information to the contrary, it seems more probable that the small diabase is younger than the granophyre; the absence of hybridization effects suggests this also.

The larger xenolith of coarse diabase in the granophyre is of such general shape as to suggest that it might have been removed from the indented part of the margin seen at the upper right in the figure. If such were the case, the xenolith has been moved to the left about 4 feet, although it still lies with its long edges nearly parallel to the contact. The smaller fragment is more irregular in shape although its sides are fairly straight and nearly parallel. At the contact with this small block the granophyre contains many sinuous, radially arranged fractures which may be due to local cooling effects brought about by the immersion of the xenolith in the granophyric magma.

DESCRIPTIVE PETROLOGY

QUARTZITE

The regularity with which massive siliceous beds alternate with relatively thin micaceous ones has been noted. By far the greater part of the formation is composed of the massive, buff, highly quartzose rock.

These siliceous beds consist of quartz, oligoclase, hornblende, diopside, and chlorite with subordinate biotite. The accessory minerals are apatite, zircon, epidote, carbonate, magnetite, pyrite, and occasional small grains of garnet. Quartz and feldspar are usually the most abundant minerals and form a uniform, fine-grained, granoblastic aggregate. When feldspar is abundant, the rock is turbid in thin section, owing to the presence of abundant, minute, dusty inclusions. The feldspar is only very slightly sericitized. The granoblastic structure of the rock is accentuated by the interstitial relationship of shreds of hornblende and chlorite which commonly lie at the contacts of the quartz and feldspar grains.

Hornblende usually occurs in small, very ragged grains. Occasionally somewhat larger porphyroblasts form diablastic intergrowths with the quartz and feldspar of the groundmass. The hornblende is pale green with faint pleochroism and low birefringence. The angle $Z \wedge c = 18° - 20°$. This mineral is frequently replaced by chlorite and carbonate.

Diopside is often abundant and may exceed hornblende in amount. It usually occurs as small, scattered, nearly colorless, xenoblastic grains and aggregates. Brown biotite is usually associated with the diopside. Although some of the biotite may be derived from the alteration of the pyroxene, most of it appears to be a recrystallization product. Occasionally large porphyroblasts of biotite are seen, which, like the hornblende, are usually poikiloblastic with respect to quartz and feldspar. Large porphyroblasts of diopside are rare but are nearly always discrete crystals without included portions of the groundmass.

In portions of the formation less affected by contact metamorphism, the original clastic structure may be preserved in part. In thin section, however, there are only rare suggestions of an original lamination. Such faintly foliated structures as are seen are due to the alternation of quartz-rich bands and somewhat wider bands of the normal quartz-feldspar-hornblende-diopside aggregate.

The massive quartzite appears to have been originally an impure, perhaps somewhat feldspathic sandstone of uniform grain and composition. The presence of abundant hornblende, carbonate, and epidote especially suggests that originally there might have been a substantial amount of calcareous matter present. It is very difficult to estimate the relative amount of material derived from the adjacent gabbro. The magnesium entering into the formation of the abundant diopside was most probably introduced in large part, inasmuch as diopside is conspicuously more abundant in the contact facies of the rock, and, moreover, may frequently be associated with introduced quartz and andesine. The presence of much apatite and, locally, of aggregates of coarse-grained magnetite in the rock at and near contacts seems to indicate their derivation from the gabbro.

The micaceous laminae are composed of an extremely fine grained aggregate of quartz, red-brown biotite, and small granules of diopside. Recrystallized feldspar is subordinate or lacking in these members. The quartz, biotite, and the usually somewhat less abundant diopside of these laminae form a fine-grained granoblastic aggregate.

A study of thin sections shows that materials derived from the gabbro has been bodily injected into the micaceous beds in some places. Stringers and small lenses of coarsely crystalline allotriomorphic intergrowths of quartz, andesine, diopside, and biotite are found. These seem to have been injected from the quartz-rich border facies of the gabbro, with some slight concomitant replacement of the body of the micaceous rock. Such injection was not noted in the massive quartzite, so that original differences in structure and composition have been accentuated by subsequent contrasts in the behavior of the two lithologies under intrusive conditions.

On the whole, cataclastic effects such as crushing and mylonitization are inconspicuous, probably because such previous events in the history of the formation are largely masked by complete recrystallization.

Most specimens examined in thin section were necessarily taken less than 75 feet from the contact of the gabbro and quartzite. The intrusion of the gabbro has, broadly speaking, led to the introduction of intermediate plagioclase, granular diopsidic pyroxene, quartz, and some red-brown flakes of biotite. Accompanying the introduction of these minerals there has been considerable reorganization of the original constituents of the rock.

DIKE ROCKS

In this paper no attempt has been made to give a detailed petrographic description of the varieties of multiple dikes. It is felt that, in view of the lithologic range and of the difficulty in determining age relations in some complexes, and in view of the highly modified condition of many of the dikes, only those interpretations supported by chemical data would be significant. However, a few of the outstanding thin-section characteristics of the abundant types will be briefly noted.

The strongly porphyritic augite lamprophyres are, almost without exception, highly modified. In some instances the original groundmass of these rocks has been completely destroyed. Most of the lamprophyres, including the hornblendic ones of more camptonitic character, now have a groundmass consisting of unaltered andesine, quartz, zoisite, granules of magnetite, carbonate, red-brown biotite, and diopsidic pyroxene. The feldspar and pyroxene form highly equigranular intergrowths, and hence the originally more allotriomorphic character of the rock is destroyed. In less severely altered lamprophyres, residual areas of turbid, zoisitized plagioclase and irregularly dispersed prismatic hornblende and altered

augite are seen. The most persistent property of the original rock is the porphyritic structure which is carried out by the disposition of the partially uralitized augite phenocrysts. The latter are frequently twinned and may show strong hour-glass structure.

Many structural varieties of diabase are found. Some are highly ophitic, but others exhibit the more characteristic interstitial development of augite with respect to plagioclase. There are varieties in which the plagioclases are definitely acicular and have a plumose disposition. In the porphyritic types either labradorite or augite alone forms the phenocrysts, and only rarely are the two associated as phenocrysts in the same rock.

Dependent largely on original differences in composition and proximity to the gabbro, two principal types of alteration products have been developed in the diabases. Some, usually the more fine grained, have been nearly completely saussuritized with the preservation of only occasional remnants of the original mesostasis. In others, in which saussuritic materials are not abundant, small flakes of red-brown biotite, probably derived in part from introduced material and in part from the original constituents of the rock, are very abundant.

The quartz porphyries, granophyres, and bostonites show no outstanding peculiarities of structure or modification. In general, albitization has been prevalent, and evidences of late-stage attack by sodic feldspar is seen in nearly all dikes of this type.

GABBRO

The most widespread and typical phase of the Cape Neddick gabbro is a uniformly coarse grained rock with a hypidiomorphic granular structure. The rock is composed essentially of labradorite, augite, biotite, hornblende, and occasionally small amounts of olivine. Accessory magnetite and apatite occur, together with additional, unusually large amounts of these same two minerals believed to be of late magmatic origin. The labradorite occurs in large, perfectly fresh crystals, many exceeding 3 mm. in length. It is commonly twinned in a very complicated manner on the Carlsbad, albite, and pericline laws. The augite, usually the most abundant ferromagnesian constituent, is pale violet and has a maximum extinction angle $Z \wedge c = 48°$. Although the rock as a whole has a hypidiomorphic granular structure, the augite is sometimes subophitic with respect to labradorite, and the individual crystals of pyroxene may then have angular outlines with sharp salients and re-entrants. More rarely large crystals of augite may have one or two small crystals of labradorite poikilophitically included within them. In some orientations the augite shows a delicate schillerization effect due to the presence of oriented metallic inclusions.

The deep-brown hornblende has a strong pleochroism with the following formula: X, pale yellow brown; Y, deep brown; Z, yellow brown. The maximum extinction angle $Z \wedge c = 18°$. This mineral occurs as unit grains attaining large dimensions and is frequently associated with augite. In the typical gabbro, hornblende is less abundant than biotite, which is red-brown, has strong absorption, and, like the hornblende, may also be in close association with augite and also with the abundant magnetite.

The structural relationships in thin section indicate that the paragenesis of the more abundant ferromagnesian minerals has been (1) augite, (2) hornblende, (3) biotite. These three are sometimes so intergrown and distributed as to suggest that the pyroxene has often controlled the distribution of the hornblende and biotite. Many rounded and irregularly embayed residuals of augite are found in the hornblende. The structural evidence concerning the relative ages of the hornblende and biotite is in general not quite so clear, although the biotite seems to be later. Biotite is likewise intergrown with augite in places though not to any such extent as the hornblende. These facts suggest a recapitulation of a part of a discontinuous reaction series. However, coarse-grained magnetite, perhaps ilmenite in part, apatite, and calcite are very abundant and far exceed in quantity the more euhedral accessory apatite and magnetite of an unquestionably orthotectic period of development. There is structural proof that the coarsely crystalline magnetite and apatite have replaced the other minerals in the rock and more especially the femic ones. Probably much of the hornblende and biotite have been developed during a late magmatic stage concomitantly with the crystallization of the magnetite and apatite.

In the anorthositic phase of the gabbro, hornblende and biotite are lacking, although augite is always present. There is little, if any, increase in the basicity of the feldspar in this phase. A very melanocratic cortlandite-like variety also occurs. This is an olivine-rich differentiate carrying large (4 mm.) crystals of this mineral and clusters of smaller euhedra, giving the rock a crudely glomeroporphyritic aspect in thin section. Usually the olivine has undergone only incipient alteration, and only the coarser partings show the presence of any feebly refracting serpentinous matter. This phase is also very rich in magnetite and carries considerable apatite as well. Although the olivine often contains small euhedra of accessory magnetite, it is also replaced by additional coarsely granular magnetite just as is the normal rock. In the cortlandite-like phase olivine is only rarely enclosed by hornblende. Augite and hornblende are quite commonly associated poikilitically, in which cases the hornblende surrounds residuals of augite. Thus we have a rock closely approaching the cortlandite type in structure rather than in the typical

association of minerals. For this reason the term cortlandite cannot properly be used without reservation.

Lithologically the dikes cutting the gabbro range in composition from diorite and quartz diorite-porphyrites to granite-aplite and even to alaskite-aplite. The most melanocratic hornblende-rich types often have a striking trachytic structure obvious even in hand specimens. All except the aplitic dikes are strongly porphyritic. They are uniformly fresh when seen under the microscope, a condition in marked contrast to that observed in the pre-gabbro dikes of the quartzite. Pegmatitic segregations in the gabbro are coarse-grained derivatives consisting essentially of hornblende, intermediate plagioclase, and carbonate.

GRANITE

The primary minerals in the granite are microcline, soda microcline, cryptoperthite, albite, oligoclase, orthoclase, quartz, riebeckite, and aegirinaugite. Accessory magnetite, zircon, apatite, sphene, rutile, and cassiterite occur together with an unusual amount of fluorite. Albite, oligoclase, and orthoclase constitute only a subordinate part of the feldspar content of the rock. Quartz, though abundant, occupies the interstices of the rock fabric. The structure of the rock is hypidiomorphic granular, and the texture is rather coarse, as many of the feldspar crystals exceed 3 mm. in length. Riebeckite is the most abundant ferromagnesian constituent in what is here considered the normal rock, and it is always associated with some aegirinaugite, with which it is often intergrown. In some places the riebeckite replaces the aegirinaugite along the (110) cleavages.

The riebeckite has intense absorption which makes definite determinations of its optical properties difficult. The extinction angle $X \wedge c = 5°$. The pleochroism varies as follows: X, dark greenish blue; Y, yellow green; Z, blue black. In sections parallel to (100) the riebeckite often shows a mottled distribution of light-blue patches which traverse the crystal apparently without any regard to cleavage.

Aegirinaugite-rich phases of the granite are sometimes found in which riebeckite occurs in diminished amount. In this type there are abundant crystals of accessory cassiterite with cross sections as large as 1 mm. The cassiterite is light brown and strongly zoned. The aegirinaugite is pleochroic in shades of green and yellow, has a high birefringence, and $Z \wedge c = 83°$ (maximum). The pleochroism varies as follows: X, olive green; Y, olive green; Z, light yellow green.

In each thin section of the granite there are usually several places where the ferromagnesian minerals have been highly altered. Such areas now consist of an intimately mixed, fine-grained aggregate of quartz, albite, rutile, leucoxene, hematite, epidote, carbonate, and very abundant fluorite.

Aside from the abundant blotchy purple fluorite in these alteration aggregates, individual crystals of fluorite are sometimes seen in the interstices of the rock. Such crystals are colorless, but in diminished light their distinct octahedral cleavage and negative relief are conspicuous. Small blebs of albite in the alteration aggregates are sometimes in optical continuity with deuteric albite which replaces the orthotectic feldspar. More rarely individual microlitic crystals of albite are sometimes developed in these areas. The structure of the aggregate is then in greater contrast with the hypidiomorphic structure of the rock as a whole.

Albitization of the feldspars by deuteric processes has been widespread. The albite commonly replaces along irregular channelways which widen at their intersections to larger blotches. Twinning effects originally present in the orthotectic feldspar are appreciably softened and may be completely obliterated. Individual crystals are often surrounded by a rim of albite where the contacts between mineral grains have served as passages of access for the late-stage fluids. Where albitization has proceeded to an advanced stage the deuteric albite itself may be polysynthetically twinned, so that contrasted types of twinning of diverse optical orientation may be found in the same crystal.

The primary essential minerals of the riebeckite microgranite dikes are the same as those of the riebeckite granite, but the proportions differ somewhat. Quartz is always more abundant in the dikes than in the typical granite and outranks all the feldspars. In the hand specimen these microgranites are moderately coarse grained, only slightly porphyritic, light gray, and of apparently uniform texture. The riebeckite is conspicuous because of its dark blue color, its prismatic development, and the size (many greater than 4 mm.) and crude parallelism of the elongate crystals. The marginal portions of the dikes are distinguished by concentrations of riebeckite.

Microscopically the riebeckite microgranites are medium grained, allotriomorphic, and only very slightly porphyritic. Sometimes the larger prismatic crystals of riebeckite have a rough flow distribution. Large crystals of quartz and feldspar are intermixed with hosts of crystals of greatly diminished size although still in interlocking relationship. There is, therefore, an extreme variability in texture from one part of a section to another, and such hiatal fabric relationships are highly characteristic.

Elongate crystals of riebeckite are interspersed with smaller anhedra of riebeckite and aegirinaugite. The latter is almost always subordinate to the riebeckite. Magnetite and apatite are the only significant accessory minerals. The distinctive alteration seen in the granite is lacking in these derivatives. The only alteration products are patches of coarse sericitic mica usually derived only from microcline.

The most characteristic feature of the microgranites is late-stage quartz flooding. The feldspars, especially microcline, have first been albitized by deuteric processes. They have then been replaced by quartz which certainly must have been relatively abundant. Individual crystals of feldspar are nearly everywhere penetrated and honeycombed by quartz. The apparently discrete patches of quartz in the albitized feldspars are in uniform optical orientation throughout the host crystal, and often with the finer-textured portions of the rock as well. Flooding with quartz was observed only in the microgranites and in one relatively quartz-rich border phase of the granite. Possibly the inequigranular fabric of these dikes can be ascribed, at least in part, to this flooding, which, it is believed, could have produced the relatively fine textured aggregates of quartz and feldspar which apparently crystallized somewhat after the development of the coarser quartz-feldspar-riebeckite and aegirinaugite of orthotectic origin.

The inequigranular structure, relative fineness of grain, and prismatic development of riebeckite of the microgranites are in strong contrast with the normal host granite. The relative scarcity of accessory minerals such as are common to the granite, the absence of intensive alteration effects in the ferromagnesian minerals, the abundance of quartz, and the nearly universal replacement of the feldspars by late-stage emanations are additional features characteristic of the microgranites as contrasted with the riebeckite granite.

SUMMARY

A study of the multiple dikes cutting the Kittery quartzite at Cape Neddick shows that they exhibit constant structural and lithologic characteristics which indicate an unusually complex history of intrusion.

The majority of dikes, both solitary and multiple, have a nearly vertical dip and strike between N 40° E and N 45° E. Widespread faulting has played a prominent part in the deformation of the dikes and dike complexes. The most conspicuous and numerous displacements are at a large angle to the regional strike of the dikes. None of the complexes is folded, and all folding of the quartzite has preceded the intrusion of the dike swarm.

Two conspicuous patterns of structural behavior characteristic only of the multiple dikes have been detected. Braiding of the components of multiple dikes is widespread. Intrasectate and extrasectate combination, and frequent reversals of such relationships, are perhaps the most outstanding structural aspects of the multiple dikes.

The lithological and structural complexity and the density of aggregation of both solitary and multiple intrusions increase with proximity to the gabbro. The most plausible reason for this is that the site of this

stock has been a focal area during several periods of intrusion prior to the emplacement of the gabbro.

Internal solidification effects in the dikes are not conspicuous, although flow banding, ocellar structure, vesiculation, and hybridization are found. Modification of the dikes by regional and contact metamorphism has obliterated many original differences which might otherwise be of diagnostic value.

The dikes in the quartzite show considerable lithologic variety. They range from olivine basalt through diabase (numerically the most abundant type), augite lamprophyre, augite andesite, diorite and quartz diorite-porphyrite to quartz porphyry, granophyre, bostonite, and aplite. Paisanitic rocks, monchiquite, and other dikes with sodic affinities related to the riebeckite granite magma are rare.

Both numerically and in bulk development basic dikes are in overwhelming abundance. Only a few granophyres, quartz porphyries, and dikes of such relatively acid composition occur. The majority of the multiple complexes consist entirely of basic members, although a few are somewhat more diversified lithologically.

Complexity due to parallelism, crushing, faulting, and contact metamorphism makes impossible the separation of individual dikes in the multiple complexes into groups representative of sharply demarcated periods of intrusion. A study of the region as a unit indicates, however, that the earliest dikes were predominantly basic. Following this basic group there is believed to have been a succession containing both acid and basic dikes. This succession would account for the granophyres, bostonites, and other acid varieties now found in multiple relationship with more basic rocks. A still later series of basic dikes, definitely transecting with respect to the earlier multiple ones, was intruded. The components of this later series did not habitually form multiple dikes.

The outstanding structural feature of the dikes as a swarm is their tendency toward parallelism. In general, their disposition is dependent upon the availability of parallel and subparallel fracture zones, but in addition these zones also had close aggregation. Localization of fracturing along the contacts of the dikes and host rock cannot be the sole explanation at Cape Neddick since intrasectate relationships are as common as extrasectate ones. Moreover, braiding is the result of frequent change from intrasectate and extrasectate relationships.

Given a structural condition in the host rock which, under the influence of regional stresses, produced a system of irregularly grouped, but in certain places densely spaced fractures, multiple dikes would still not necessarily result. The structural relationships show that frequently repeated fracturing is required to produce the observed associations and

complexity. Thus another significant factor is the repetitious nature of crustal movements, each series of which has been so resolved as to contribute to the parallelism of the fracture system. Much of the difficulty of establishing a possible cause of recurrent fracturing is due to inadequacy of exposures. Hence the major elements of regional structure and relationships to other rock units cannot be determined.

The recurrent periods of fracturing have been correlated with the intrusion of magma. The regional movements appear to have caused sympathetic movements of magma and influenced its course of differentiation. There are a few multiple dikes whose lithological association manifestly reflects a fairly well defined differentiation. It is probable that in this area repeated fracturing and concomitant emplacement of systematically differentiated magma fractions are interdependent aspects of the regional phenomena.

WORKS TO WHICH REFERENCE IS MADE

Hitchcock, C. H. (1861) *Preliminary report upon the natural history and geology of the State of Maine, General report on the geology of Maine,* p. 214.

Jackson, C. T. (1838) *Second report on the geology of the State of Maine,* Augusta, Maine, p. 72.

Katz, F. J. (1918) *Stratigraphy in southwestern Maine and southeastern New Hampshire,* U. S. Geol. Survey, Prof. Paper 108, pt. 1, p. 165-177.

Kemp, J. F. (1890) *On the dikes near Kennebunkport, Maine,* Am. Geol., vol. 5, p. 129-140.

Powers, S. (1915) *The origin of the inclusions in dikes, Part II,* Jour. Geol., vol. 23, p. 172.

Thomas, H. H. and Bailey, E. B. (1924) *Tertiary and post-Tertiary geology of Mull, Loch Aline, and Oban,* Mem. Geol. Survey Scotland, p. 356 (Portion by Bailey and Richey).

Wandke, A. (1922) *Intrusive rocks of the Portsmouth Basin, Maine and New Hampshire,* Am. Jour. Sci., 5th ser., vol. 4, p. 139-158.

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PETROGRAPHY OF THE CRYSTALLINE LIMESTONES AND QUARTZITES OF THE GRENVILLE SERIES

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ABSTRACT

The crystalline limestones of the Grenville series in the vicinity of Kingston are similar to those in other parts of eastern Ontario and in the Adirondacks. They consist of pure and dolomitic limestones. Practically all of them contain small quantities of noncarbonate minerals. Some of these are magnesian silicates which probably derive their magnesium content from the original dolomite of the limestone. Some of the noncarbonate constituents evidently were formed by additions of certain constituents from the abundant intrusives. Garnet apparently did not form commonly in the limestone because of deficiency of alumina and possibly of manganese.

Quartzites are interbedded with the limestones. Foreign material is much less abundant in them, as the quartz of the original sandstones was no doubt much more inert chemically than was the carbonate of the limestone.

INTRODUCTION

The Grenville series of Eastern Ontario and Western Quebec consists of an assemblage of highly metamorphosed sediments. The chief rock types are crystalline carbonate rocks, quartzites, and schists of various kinds. Some of the carbonate rocks are dolomitic; others are rather pure lime rocks. The two types are usually indistinguishable in hand

specimens, and the term crystalline limestone will be used in the following discussion to cover all crystalline carbonate rocks without regard to their actual composition. Only the quartzites and crystalline limestones of southeastern Ontario will be considered here.

CRYSTALLINE LIMESTONES

GENERAL STATEMENT

A detailed study of the crystalline limestones of the Haliburton-Bancroft area was made by Adams and Barlow (1910, p. 192-226). Those near Kingston were investigated by Baker (1916) and those of the County of Leeds by Baker (1922) and by Wright (1923, p. 18, 19).

The sedimentary origin of the Grenville series is certain, and in general the structure is evident. In some of the limestones there is a banding which has been interpreted as bedding. Usually it is due to concentration of small quantities of minerals, other than carbonates, in linear arrangements. It seems probable that in large part this apparent stratification is due to shearing rather than to sedimentary processes. In most places it is approximately parallel to the actual bedding as shown by the quartzite and gneiss with which the carbonate beds are interstratified.

Adams and Barlow (1910, p. 193) note the fact that dolomitic varieties are indistinguishable in general appearance from those that are true limestones. They state that in some parts of the Haliburton-Bancroft area the limestones are almost free from silicates. In other places the carbonate rocks contain considerable quantities of silicates and other non-carbonate minerals. They record a list of 37 mineral species (Adams and Barlow 1910, p. 198). Wright (1923, p. 19) suggests a contact metamorphic origin for some of the minor constituents of the limestone of the Brockville-Mallorytown area, by recrystallization of a siliceous dolomite bed and addition of some material of igneous origin.

Cushing, Fairchild, Ruedemann, and Smyth (1910) describe Grenville limestones on the south side of the St. Lawrence River. They point out that the marginal parts of some beds of limestone contain more non-carbonate minerals than occur in the central parts. This they attribute to the gradation of original limestone into shales or sandstones above or below. In some cases, however, the limestone has been modified by igneous rocks.

Cushing (1918) describes crystalline limestones in the Gouverneur area of New York State. He divides them into two groups, one of which is relatively pure, the other impure. The former is white, of medium texture, and contains only a small amount of phlogopite in tiny flakes, graphite, and a little pyrite. It is 95 to 99 per cent calcite. There are also some pure limestones that are gray or bluish gray and commonly of a

banded character. There are some dolomitic areas but these are not extensive. The impure limestones are siliceous, as they grade into quartzites, but in other places serpentine, talc, light-colored pyroxene, and scapolite are present.

Buddington (1934, p. 110-112) states that in the Hammond and Antwerp quadrangles of New York State only a small part of the crystalline limestone is relatively pure. Most of it contains disseminated silicates or nodules of silicates. Much of it is veined with quartz or is cut by dikes or lenses of granite or pegmatite. Some beds of limestone are replaced partly or completely by pyroxene, scapolite, or quartz. Some of the pegmatite veins are replacements of limestone, and all gradations occur between normal pegmatite dikes and limestone with disseminated quartz and feldspar grains. Much of the limestone contains tremolite; phlogopite is present in nearly all, and nodules in limestone contain brown tourmaline, brown scapolite, and some chondrodite. Spinel is also present in some beds.

Analyses of samples of limestone from the marble quarry at Gouverneur show the following compositions (Buddington 1934, p. 219).

Insoluble	3.55	1.26
Al_2O_3	0.13	0.65
Fe_2O_3	0.08	0.29
$MgCO_3$	6.40	7.50
$CaCO_3$	87.06	87.47
H_2O	1.68	1.46
S	0.05	0.02
	98.95	98.65

Reed (1934) states that the purest limestone of the Potsdam quadrangle, New York, contains only a small percentage of minerals other than carbonates but that these are seldom lacking. He lists as species commonly present phlogopite, graphite, serpentine, chondrodite, diopside, biotite, feldspar, and quartz.

Wilson (1924) states, with regard to the Grenville limestone of the Maniwaki area, that the banding is due in part to the distribution of pyroxene, phlogopite, and other impurities in alternating parallel zones. He considers that these zones represent original bedding but that the minerals have been formed secondarily by the action of igneous solutions, since the elements contained are not ordinarily present in sedimentary rocks.

It seemed possible that a detailed study of the mineralogy of some of the limestones, especially where they are intruded by pegmatite dikes, might throw some light on the origin of the constituent minerals other than calcite. For this purpose limestones from localities near Portland, near Morton, and near Verona north of Kingston, Ontario, were examined in some detail.

Calcite and dolomite are the most abundant minerals of the limestone. They were distinguished in thin section by the differences in the position of the twinning planes (Rogers, 1929), by optical examination of grains in pulverized specimens, and by staining tests. Grains were found with indices $N_e = 1.5059$ and $N_o = 1.6790$ and thus are undoubtedly dolomite. Specimens from a pinkish-brown weathering band contained crystals that gave indices $N_e = 1.5054$ and $N_o = 1.700$ which agree with those of magnesite. Silver nitrate and potassium chromate were used for staining. Polished surfaces were treated with silver nitrate, thoroughly washed, and immersed in a 10 per cent solution of potassium chromate. Calcite turns deep reddish brown, but dolomite is unaffected.

From the evidence that follows it seems clear that magnesium and calcium in the silicates now found in the limestones were supplied wholly from the limestone. Silica, boron, fluorine, chlorine, phosphorus, sulphur, and hydroxyl were contributed by the intrusives.

LIMESTONES NEAR PORTLAND, ONTARIO

Crystalline limestone is exposed in an old quarry and in a road cut east of the village of Portland. A pegmatite dike cuts the limestone at the quarry.

Minerals recognized in hand specimens of the limestone include brownish tourmaline, chondrodite, phlogopite, apatite, and serpentine. Thin sections of the limestone show a considerable variety of minerals and two types of texture. In one textural type the minerals present a mosaic pattern (Pl. 1, fig. 1). The carbonate grains commonly show curved or broken polysynthetic twinning bands and well-marked cleavage. The grain size varies from diameters of less than 1 millimeter to diameters of 5 and 6 millimeters. Usually the grain boundaries are smooth and unbroken, but in some specimens large grains are separated by zones of crystal fragments. A coarse-grained mosaic pattern is typical of bands of pure white limestone containing only a few scattered noncarbonate minerals. Nearer the pegmatite the limestone is finer in texture. The grains vary from 0.1 to 5 millimeters, the contacts between them are less regular, and twinning in the carbonates is less perfect. There is also a larger proportion of noncarbonate minerals. In samples from the quarry, pyroxene is next to the carbonates in abundance; in the rock from the road cut it is much less abundant. There is clearly an increase in the proportion of pyroxene as the pegmatite dike is approached. The wall rock of the dike is a pyroxenite containing a few scattered grains of carbonates (Pl. 1, fig. 2). The pyroxene is colorless and is probably diopside.

Quartz is present in nearly all thin sections. Two types occur. One

is in the form of rounded grains scattered through the carbonates; the other has slightly lower indices of refraction, lower birefringence, and shows curved fractures. No means were found of determining whether the quartz was an original constituent of the sediments or was introduced.

Thin sections show that microcline and bytownite in about equal quantities are sparingly present in limestone near pegmatite dikes. Bytownite and the second type of quartz mentioned above commonly form intergrowths which extend out into calcite grains and include rounded blebs of calcite. Similar relations have been described by Agar (1923) and interpreted as replacement of calcite by the quartz-feldspar intergrowth.

Graphite is abundant in the crystalline limestone. It lies between the carbonate grains, in some places conforming to the grain boundaries, in others cutting directly across them. All of the grains of graphite are roughly ovoid with average dimensions 2 millimeters by 0.5–1 millimeter. Graphite usually occurs with the second type of quartz.

Pyrite is fairly common in the crystalline limestone and is more abundant near the pegmatites than elsewhere. It is present in the pegmatites as euhedral grains and as tiny veinlets. In the limestone it occurs only as euhedral grains.

Pyrrhotite is less abundant than pyrite and is confined to the dikes and limestone near the dikes.

Chalcopyrite is associated very sparingly with pyrite and pyrrhotite in the pegmatite dikes. None was seen in limestone.

Sphene occurs as small wedge-shaped crystals of a deep reddish brown. Some of them show sufficient alteration to appear white by reflected light.

In addition to these minerals which were found in thin sections examined, several others are recognizable in hand specimens. These include tourmaline, chondrodite, phlogopite, apatite, and talc. Large crystals of tourmaline occur in both the pegmatites and the limestone. The mineral is reddish brown, and the indices of refraction indicate that it is the variety dravite. Similar tourmaline has been described from pegmatites and limestones in other parts of eastern Ontario (Bruce, 1916; Harcourt, 1933).

Chondrodite is present as irregular aggregates of reddish-brown to orange-brown grains erratically distributed. Thin sections show that in some of these aggregates quartz of the lower index type is present.

Phlogopite of two types was found. In one type brown crystals, the edges of which are greenish, occur with dravite. The other is pale amber and associated with apatite in pure white crystalline limestone.

Apatite occurs in pale blue to pale green barrel-shaped crystals. The usual size is 5 millimeters in length and 1 to 2 millimeters in diameter.

One small piece of talc was found in pegmatite, and a few flakes in crystalline limestone adjacent to the dike.

Serpentine is rather common, and in places the rock is an ophicalcite. Serpentine also occurs as a fracture filling.

LIMESTONES NORTH OF MORTON, ONTARIO

The rock ridge crossing the highway on the north side of Morton Creek consists of interbedded quartzite, crystalline limestone, and gneisses with *lit-par-lit* injections of granite. The beds of gneiss and quartzite stand out in relief. The crystalline limestone has weathered into hollows and is covered by reddish soil, but a road cut through part of the ridge provided fresh material for study.

The limestone is darker than that at Portland, varying from gray to dark green. In places it contains fragments of gray gneiss probably from beds broken by stresses under which the limestone flowed.

A pegmatite dike 3 feet in greatest width intrudes the limestone in the road cut. It probably is a tongue from a larger mass that outcrops on the top of the ridge farther east. The contact of the dike with the limestone is fairly sharp, but the two rocks are separated by a narrow band of pyroxene next to which the limestone is very friable (Pl. 1, fig. 5).

The grain size of the crystalline limestone is about the same as that at Portland. Most of the grains are elongated and lie with their long axes parallel.

A thin section across the contact of the pegmatite dike with the limestone shows a gradation. At the contact the carbonate has been largely replaced by quartz and diopside, and the proportion of these minerals to carbonate decreases rather sharply away from the contact. (Pl. 1, figs. 3, 4).

LIMESTONES AT BREWER'S MILLS, ONTARIO

Crystalline limestones at Brewer's Mills are intruded by a pegmatite dike and a diabase dike. The limestones are interbedded with bands of rusty-weathering gneiss approximately a foot thick; at one place the rock has a brecciated character with fragments of gneiss enclosed in limestone similar to that described at Morton.

In texture and mineralogy the limestones are similar to those previously described, but here there are masses of tremolite and pale-green diopside and a few crystals of biotite. The tremolite is in white interlocking fibres which are on the average 0.5 centimeter in length.

LIMESTONES NEAR VERONA, ONTARIO

Limestone outcrops at several localities near the village of Verona. Three of these were examined: one half a mile north of the village,

one immediately south of the village, and one at the east of Thirteen Island Lake, 7 miles northeast of the village. At the two latter places basic dikes intrude the limestone, and at Thirteen Island Lake movements later than the intrusion of the dikes have fractured them and scattered the fragments through the limestone.

The limestone outcropping east of the highway, north of Verona, is white with green mottling. It consists of white carbonates similar to those of other limestones described, and serpentine. These two minerals are approximately equal in abundance and make up the greater part of the rock. Scapolite and muscovite are present in small amounts. There is no direct evidence to show the original mineral from which the serpentine formed. It clearly replaces the carbonates, and, as the relations are similar to those between diopside and carbonates in the Portland rock, it is probable that the serpentine formed from diopside.

Portions of the rock from this locality north of Verona are pink to reddish. In thin section samples of this kind show rounded remnants of carbonate in a pink groundmass of carbonate of different type. Staining tests on polished surfaces show that the rounded remnants are calcite. Indices of refraction determined on different fragments of the matrix range from those of dolomite to those of magnesite.

Most of the limestone on the shore of Thirteen Island Lake is nearly pure white. In places there are gray and greenish-gray types. The rock in a road cut at the end of the lake is traversed by thin, curving, dark gray lines along which plates of graphite are plentiful and, in places, tiny fragments of the fractured basic dikes. Undoubtedly the curved lines are not the traces of bedding planes but are due to flow in the limestone. Near the basic dikes the limestone contains a little biotite.

The limestone south of Verona is very similar to that at Thirteen Island Lake excepting that it is gray. It contains larger quantities of minerals other than carbonates, especially near the basic dike by which it is intruded. Species recognized are biotite, which is present in plates as much as 3 millimeters in diameter, chondrodite, apatite, diopside, chalcopyrite, pyrrhotite, and quartz.

DEVELOPMENT OF THE CRYSTALLINE LIMESTONES

GENERAL STATEMENT

It is clear that the Grenville rocks are mainly of sedimentary origin. They were originally an interbedded series of sands, clays, and limestones which are now quartzites, gneisses, and crystalline limestones. Some of the crystalline limestones are dolomitic, and others are relatively pure lime rocks. Both types contain minor amounts of minerals which do not occur in unmetamorphosed carbonate rocks. The problem of determin-

ing the origin of various constituents of the limestones is complicated by the fact that evidently the whole series has passed through orogenic disturbances, which must have profoundly affected the various rock types. There also has been invasion of these sediments by igneous rocks of at least two periods. It is a matter then of attempting to decide which constituents have been formed directly by reorganization and recrystallization of material originally contained in the limestone and which have been derived in whole or in part from material contributed by the igneous rocks. The problem would be simpler if it were possible to trace the progress of contact metamorphism from the intrusive rock through to unaltered limestone. In the present study no limestone has been found that can be considered unmetamorphosed; it is likely that none exists in the Grenville Series (Bain, 1923). Reference has been made to little metamorphosed limestone in the Haliburton-Bancroft area (Adams and Barlow, 1910, p. 192), but it may be doubted if any limestone of undisputed Grenville age has escaped intense metamorphism.

ORIGIN OF CARBONATE CONSTITUENTS

Calcium carbonate is the normal original constituent of limestone. Its recrystallization into the calcite of crystalline limestone takes place in the ordinary course of dynamic metamorphism or of contact metamorphism where the containing pressures are sufficient to prevent the dissociation of the carbonate.

The presence of magnesium carbonate either as dolomite or as magnesite raises the problem of the origin of dolomite and magnesite. Since, however, dolomite occurs commonly in practically undisturbed rocks it may be assumed that the Grenville limestones which are dolomitic, were so before the disturbances and before intrusions to which their crystalline character may be attributed. This conclusion is further corroborated by the lack of any apparent relation between the proportion of magnesium in the rock and its proximity to intrusives. Nonmagnesian limestones occur in contact with both granite and basic dikes.

ORIGIN OF NONCARBONATE CONSTITUENTS

The minor minerals in the crystalline limestones may be grouped as follows:

(1) Silicates, which are simply compounds of silica, calcium, and magnesium—diopside, talc.
(2) Minerals which contain elements such as fluorine or chlorine, hydroxyl, sulphur, phosphorus—apatite, dravite, chondrodite, phlogopite, scapolite, and sulphides.
(3) Graphite.

It is significant that so little garnet is developed in the limestone. Miller and Knight (1914) record the occurrence of a garnet-epidote-hornblende rock east of Gilmour. They consider it to be a limestone bed altered by granite. Possibly at that place the limestone contained impurities not elsewhere abundant, as it is noted that at contacts of the blue limestone and granite there is little or no alteration.

Garnet, which occurs rather commonly in the gneisses, contains some manganese. The presence of garnet in one bed of the series and the absence of that mineral in most other beds seem probably to be due to lack of necessary material for garnet formation in the latter cases. It may have been lack of some essential constituent or it may have been the lack of manganese in the limestone beds. Presence of that element might permit the crystallization of garnet in the gneisses at a temperature below that necessary to form garnet from the material available in the limestone. Since nonaluminous silicates such as diopside are found in the limestone, it seems likely that lack of alumina was the factor which prevented the formation of garnet. Limestones normally have little alumina (Clarke, 1924), and it follows that alumina was not introduced by the intrusives except locally as indicated by the contact zone at Gilmour. In the analyses referred to there is usually a definite ratio between alumina and iron and silica. Hence the formation of diopside shows that silica, in addition to the small amounts originally present, must have come from the intrusives.

The question of the source of the magnesium in the minor minerals of the first group is important. Clearly, if the unmetamorphosed original of the crystalline Grenville was magnesian in places, as has been assumed, there was abundant magnesium to combine with any silica introduced by the intrusives. Different modes of origin of the magnesite lenses that occur in the Grenville of Argenteuil County have been proposed by various authors. Wilson (1917, p. 33) concludes that they

> "have been formed by the replacement of the limestone member of the Grenville series through the agency of magnesia rich solutions",

and he suggests that the source of the solutions was the pyroxene gabbro, pyroxene diorite, and pyroxene syenite series of intrusives. Bain (1924) also attributes the deposits to reactions between magnesia-rich solutions and limestone.

On the other hand, Wilson (1926) concludes that in the formation of talc in the Madoc area original dolomitic limestone was altered to talc through the agency of hot siliceous waters from neighboring intrusions of granite. This reaction is expressed in the following equation:

$$3(MgCO_3\ CaCO_3) + 4SiO_2 + H_2O = 3MgO\ 4SiO_2\ H_2O + 3CaCO_3 + 3CO_2$$

If the dolomitic character of the limestone be an original feature it seems reasonable to believe that the serpentine and diopside which occur in the limestone may have derived their magnesia from original dolomite.

Minerals such as apatite, dravite, chondrodite, and scapolite contain phosphorus, fluorine, boron, hydroxyl, sulphur, and chlorine. These elements are doubtless present in magmas and so may be given off in the pegmatitic stage under suitable conditions. It is probable that the small isolated crystals of the minerals listed above were formed by reactions between the calcium and magnesium carbonates and these volatile elements, passing out into the limestones as gases from the cooling igneous rocks.

Buddington (1934, p. 110-112) concluded that volatile elements such as chlorine, fluorine, and to a minor extent phosphorus, were introduced into the limestones to form noncarbonate constituents of the limestone. Reed (1934, p. 57) considered that the silicates and quartz are due partly to recrystallization of original material in the limestone and partly to introduction from external sources.

Graphite occurs in some of the limestones. It is difficult to offer a satisfactory explanation of its origin. Wilson (1919) has shown rather conclusively that it replaces the calcite of the limestone. Graphite occurs in the limestone both at Brewer's Mills and at Morton. It is also present in the pegmatites at those places, but there is no conclusive evidence that the limestones have derived graphite from the dikes.

QUARTZITES

Quartzite probably makes up a smaller proportion of the Grenville series than does limestone but owing to the great difference in resistance to weathering it is much more widely exposed.

Wright's (1923, p. 14-15) study of the quartzites of the Brockville-Mallorytown area shows a considerable variety in the rocks he includes under that name. He tabulates the results of a number of microscopic analyses.

He states that in the typical quartzite the feldspars are disseminated throughout the rock and have been granulated in the same way as the quartz but that near the granite the feldspars occur in bunches and in relations which he attributes to impregnation of the quartzite by solutions from the granite. As will be seen from the table, some of the quartzite contains abundant light-green diopside.

Wright states that it "probably originated through the recrystallization of calcium, magnesium, and quartz in the original sandstone."

In one specimen listed above there is 15 per cent diopside, which seems a large amount to have formed in that way. In view of the clearly contact

metamorphic character of diopside in limestone it seems more likely that diopside occurring in quartzite has been derived by silicification of carbonate impurities in the quartzites by siliceous solutions from the granite magma.

TABLE 1—*Mineral composition of Grenville quartzite*

Specimen Numbers	49	382	9	61	56	62	5	50	13
Quartz	98.2	92.1	89.5	75.0	75.0	71.0	54.4	52.0	41.9
Orthoclase	.....	.....	0.7	9.8	11.8	11.9	.....	.....	.....
Microcline	.....	2.2	.....	.....	.....	.....	5.7	27.2	21.0
Oligoclase-Albite	.....	1.5	2.9	12.0	10.0	9.5	19.7	12.1	27.0
Biotite	0.6	2.3	1.4	0.3	0.5	0.8	.....	2.5	4.7
Diopside	.....	.....	0.8	1.5	.....	.....	15.0	.....	.....
Magnetite	0.4	0.9	2.1	0.6	0.9	1.1	0.8	1.0	1.2
Tourmaline	.....	0.2	0.3	.....	0.3	.....	0.6	.....	3.1
Sphene	.....	.....	0.5	0.6	1.9	.....	1.2	.....	0.6
Calcite	.....	.....	.....	.....	.....	5.4	2.1	2.3	.....
Accessories	0.8	0.7	2.4	.....	0.6	0.3	0.5	2.9	0.5
Total	100.0	99.9	100.6	99.8	101.0	100.0	100.0	100.0	100.0

49—Along road half a mile east of Junetown.
382—One mile northeast of Brockville. (Collected by M. E. Wilson.)
9—South of Grand Trunk Railway (Canadian National) 2 miles west of Lyn station.
61—Doctor Island, Little Grenadier group.
56—Along road half a mile north of Rockfield.
62—Lot 5, south side of Hill Island.
5—Along brook one mile south and east of Lyn.
50—East of road from Mallorytown to Pooles Resort and one mile from Pooles Resort.
13—South of Grand Trunk (Canadian National) 4 miles west of Lyn station; Nos. 50 and 13 taken within 4 inches of granite contact.

The minor minerals listed by Wright—biotite, magnetite, tourmaline, apatite, zircon, pyrite, and sphene—may have had somewhat different histories. Sphene he considers to have been residual material as the crystals are rounded and corroded. Magnetite also may be recrystallized original material. Tourmaline and biotite possibly formed by the action of emanations from the intrusive reacting with original material in the sediments.

A rounded mound in the center of the inlier of pre-Cambrian rocks in the valley at Jackson's Mills consists of white vitreous quartzite so massive that bedding is recognizable only at a few places. Under the microscope the rock is seen to consist largely of quartz occurring as grains of two sizes. Larger ones are surrounded by smaller ones, but in both cases the grains form a closely interlocking mosaic. In places the mosaic

contains a little microcline and a very little sericite. Cubes of pyrite occur sparingly. Surrounding the grains of the mosaic near pyrite there is a brown stain no doubt the result of weathering of the iron sulphide.

A partial analysis of this rock gave the following result:

SiO_2	95.35
Fe_2O_3-Al_2O_3	2.32
Other constituents by difference	2.33
	100.00

The iron in the analysis is no doubt present as pyrite and iron hydroxide. The alumina is largely in the microcline. Other constituents include K_2O, H_2O, and very small quantities of zirconium and possibly some other oxides.

At Morton, quartzite is interbedded with limestones. The beds are 3 to 4 feet thick. They produce a corrugated surface along the top of the ridge where they weather slightly in relief. The rock is very similar to that just described. A thin section showed that zircon occurs in very small quantities.

CONCLUSIONS

From the facts stated above it is concluded:

(1) The general absence of garnet from the crystalline limestones shows that little alumina and iron were present in the original limestone and that neither was added by the intrusives.

(2) Silica was added by the intrusives and resulted in the formation of nonaluminous silicates of calcium and magnesium from the calcium and magnesium carbonates of the original rock.

(3) Minor silicates scattered through the crystalline limestone were formed by reactions between gaseous emanations from the granite or pegmatites and the original carbonates. The widespread occurrence of these minor constituents is explained by the fact that the limestone is everywhere intimately intruded by igneous masses.

(4) Quartzite of the Grenville series presents a much simpler assemblage of minerals than does the limestone, and, as conditions must have been the same for both rocks this greater simplicity must have been due to the character of the original constituents.

(5) Quartz in the quartzites did not react with the emanations from the intrusive to any marked extent. Some of it may be due to introduction of silica, but, if so, quartz from that source is now indistinguishable from original quartz grains.

(6) Microcline in the quartzite may be the result of recrystallization of original clay containing some absorbed potash.

(7) The sparsely disseminated pyrite in the quartzite was very likely

the result of introduction of iron and sulphur rather than the reaction of sulphur on original iron-bearing minerals. Limonite is clearly due to near surface oxidation of pyrite.

WORKS TO WHICH REFERENCE IS MADE

Adams, F. D., and Barlow, A. E. (1910) *Haliburton-Bancroft Area,* Geol. Survey Canada, Mem. 6, p. 192-226.

Agar, W. M. (1923) *Contact metamorphism in the western Adirondacks,* Am. Philos. Soc., Pr., vol. 62, p. 113.

Bain, G. W. (1923) *Almandite and its significance in the contact zone of the Grenville limestone,* Jour. Geol., vol. 31, p. 650-668.

——— (1924) *Types of magnesite deposits and their origin,* Econ. Geol., vol. 19, p. 425.

Baker, M. B. (1916) *Geology of Kingston and vicinity,* Ontario Bur. Mines, Ann. Rept., vol. 25, pt. 3, p. 5.

——— (1922) *Geology and minerals of the county of Leeds,* Ontario Dept. Mines, vol. 31, pt. 6, p. 4.

Bruce, E. L. (1916) *Magnesian tourmaline from Renfrew, Ontario,* Mineral. Mag., vol. 18, p. 133-135.

Buddington, A. F. (1934) *Geology and mineral resources of the Hammond, Antwerp, and Lowville quadrangles,* N. Y. State Mus., Bull. 296, p. 110-112, 219.

Clarke, F. W. (1924) *Data of geochemistry,* U. S. Geol. Survey, Bull. 770, 5th ed., p. 564, 578.

Cushing, H. P. (1925) *Geology of the Gouverneur quadrangle,* N. Y. State Mus., Bull. 259, p. 18-22.

———, Fairchild, H. L., Ruedemann, Rudolf, and Smyth, C. H., Jr. (1910) *Geology of the Thousand Island region,* N. Y. State Mus., Bull. 145, p. 29.

Harcourt, G. A. (1933) *Brown tourmaline from Frontenac and Renfrew counties, Ontario,* Am. Mineral., vol. 18, p. 356-358.

Miller, W. G., and Knight, C. W. (1914) *Pre-Cambrian geology of southeastern Ontario,* Ontario Bur. Mines, vol. 22, pt. 2, p. 85.

Reed, J. C. (1934) *Geology of the Potsdam quadrangle,* N. Y. State Mus., Bull. 297, p. 12-13, p. 57.

Rogers, A. F. (1929) *Polysynthetic twinning in dolomite,* Am. Mineral., vol. 14, p. 245-250.

Wilson, M. E. (1917) *Magnesite deposits of Grenville district, Argenteuil County, Quebec,* Geol. Survey Canada, Mem. 98, p. 30-34.

——— (1919) *Mineral deposits in the Ottawa Valley,* Geol. Survey Canada, Summ. Rept., pt. E, p. 35, 40.

——— (1924) *Arnprior, Quyon, and Maniwaki areas, Ontario and Quebec,* Geol. Survey Canada, Mem. 136, p. 20.

——— (1926) *Talc deposits of Canada,* Geol. Survey Canada, Econ. Geol. ser. 2, p. 27, 28.

Wright, J. F. (1923) *Brockville-Mallorytown map area,* Geol. Survey Canada, Mem. 134, p. 14-19.

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Plate 1

Photomicrographs of Limestones

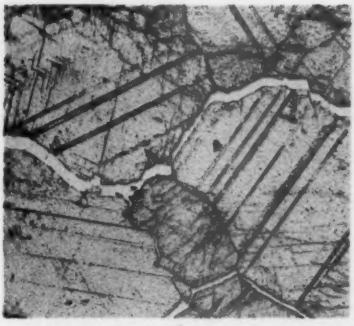
Figure 1.—Mosaic Texture in Limestone.
Portland, Ontario, one nicol (×40).

Figure 2.—Carbonate Partially Replaced by Pyroxene.
Portland, Ontario, crossed nicols (×72).

Figure 3.—Diopside Reaction Rim between Carbonate and Microperthite.
Morton, Ontario, crossed nicols (×72).

Figure 4.—Diopside Replacing Microcline.
Reaction rim around a remnant of carbonate, Morton, Ontario, crossed nicols (×72).

Figure 5.—Diopside Reaction Rim.
Between microperthite and carbonate. Large quartz grains in diopside. Upper half is pegmatite, lower half limestone, Morton, Ontario, crossed nicols (×72).



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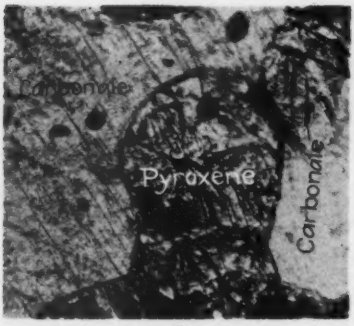

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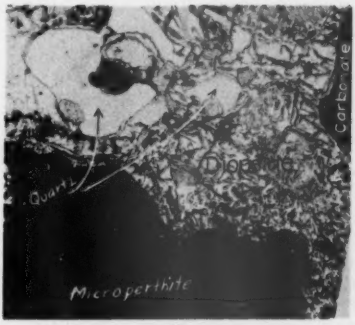

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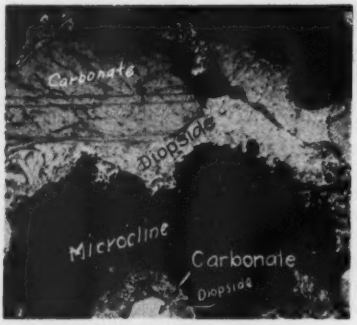

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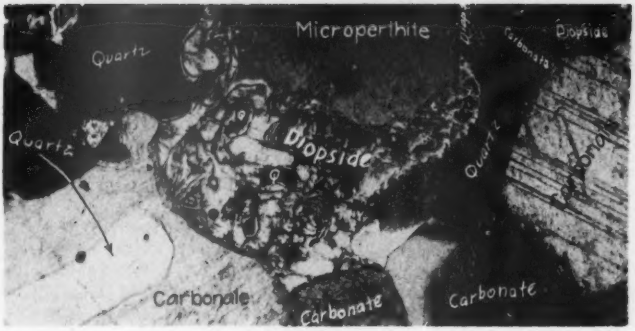

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PHOTOMICROGRAPHS OF LIMESTONE

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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WISCONSIN SILURIAN BIOHERMS (ORGANIC REEFS)

BY ROBERT R. SHROCK

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ABSTRACT

The Silurian of Wisconsin contains biohermal structures at several stratigraphic horizons. These bioherms are essentially cuboidal, domal, or ridgelike masses of porous and cavernous, crudely bedded and poorly jointed, barren to highly fossiliferous dolomites which grade peripherally outward and vertically upward and downward into well-bedded, dense to saccharoidal, relatively unfossiliferous dolomites. The approximately contemporaneous surrounding strata, which commonly are thin bedded, first dip away from the reefy mounds at relatively low angles, except where differential slumping has steepened the initial dip, and then become essentially horizontal and take on the lithological character of the normal stratigraphic sequence of the region.

The complexity of the fauna on some of the bioherms shows clearly that no single phylum of organisms played the dominant role in the construction. On the contrary, it is usually obvious that many different classes of organisms made large and important contributions to the growing calcareous mounds. Chief among the bioherm builders were algae, hydrozooid stromatoporoids, tabulate corals, camerate crinoids, cryptostomatous bryozoans, protrematous and telotrematous brachiopods, pelecypods, gastropods and cephalopods, and trilobites.

Biohermal sites, especially at Wauwatosa, Milwaukee, and in the vicinity of Cedarburg and Grafton, were the main sources of the several large Silurian collections made by local amateur collectors in the closing decades of the nineteenth century; they still remain among the best collecting sites. The interbiohermal strata may or may not be fossiliferous, but if they are their fauna tends to be small and somewhat different from the typical biohermal assemblage.

The bioherms appear to have accumulated in fairly shallow water and to have stood as low mounds a few feet to a few tens of feet above the surrounding bottoms where fine calcareous muds were being deposited.

INTRODUCTION

The fossiliferous "coral reefs" in the Middle Silurian (Niagaran) strata of eastern Wisconsin (Fig. 1) have been known for over 75 years and have been referred to frequently since their discovery and correct interpretation by James Hall in 1862. It was not until 1877, however, when T. C. Chamberlin described them in detail, that they became well known. Since the latter date most references to the Wisconsin Silurian have dealt with the stratigraphic relations and correlation of the reef-bearing strata. None so far has dealt specifically with the bioherms in as much detail as Chamberlin's original report.

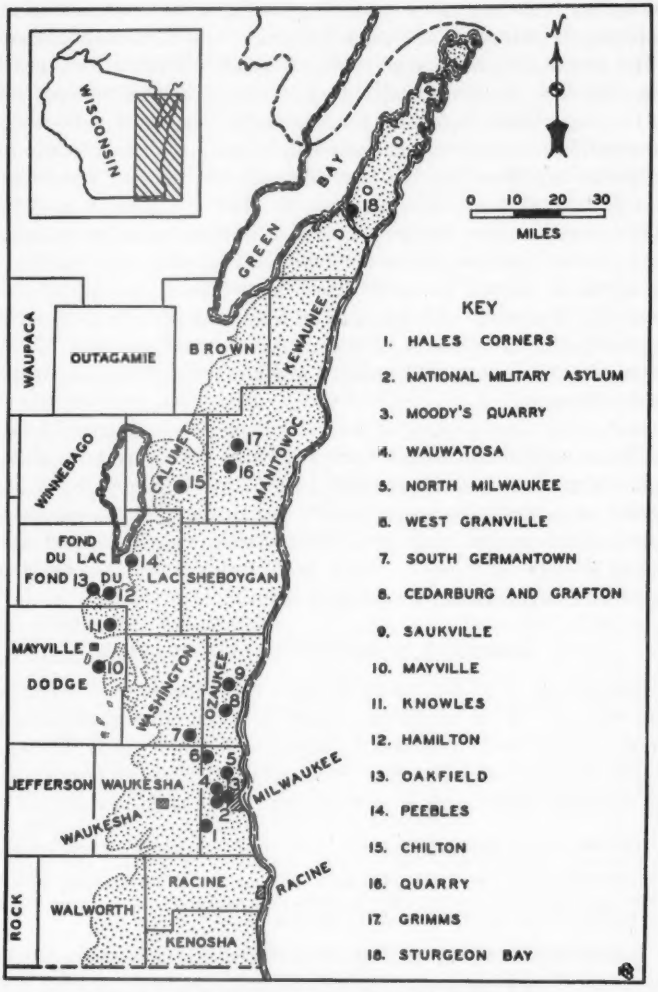


FIGURE 1.—*Index map of a part of eastern Wisconsin*
Showing area of Silurian rocks (stippled) and locations of bioherms (numbered).

Studies of the Wisconsin bioherms by the writer, at irregular intervals during the past 10 years, have yielded some new and significant data. Quarries, road cuts, and numerous other excavations, not available for study to earlier investigators, have allowed a rather thorough examina-

tion of the bioherms and surrounding strata. The time now seems appropriate for recording the more important and significant discoveries.

The so-called "coral reefs" of the Wisconsin Niagaran are nearly all true bioherms (Cumings and Shrock, 1928a, p. 599; Cumings, 1932, p. 333)—*i. e.*, organic reefs built not dominantly by corals but rather from the contributions of many animals and of certain algae. In the following discussion, therefore, the term bioherm will be used for these reefy mounds and, if the dominant fauna or flora is apparent, a qualifying adjective will indicate the important reef-building organism or organisms.

The writer has been assisted in the field at different times by C. E. Kemp, W. L. Wilgus, and J. E. Havard, students in geology at the University of Wisconsin. He has also profited greatly from field excursions and discussions with Dr. E. R. Cumings, Indiana University, Dr. W. H. Twenhofel, University of Wisconsin, and Dr. Ira A. Edwards, Milwaukee Public Museum.

Field work during parts of several summers was financed jointly by the Wisconsin Geological and Natural History Survey and the Milwaukee Public Museum. Both organizations have kindly granted the writer permission to publish this report. A ten-day field trip in the spring of 1934 was financed from a small grant from the Penrose Bequest of the Geological Society of America made to Twenhofel for the study of the Ordovician and Silurian strata of Wisconsin.

SUMMARY OF PREVIOUS INVESTIGATIONS

Hall (1862, p. 63-64) seems to have been the first to recognize the reef character of the fossiliferous mounds in the Silurian of eastern Wisconsin. He correctly sketched the stratigraphic relations of the biohermal (reef) core and interbiohermal (interreef) strata of the now well-known Schoonmaker bioherm at Wauwatosa and wrote as follows concerning it:

> "The entire mass appears like a coral reef, where the broken corals and shells are packed in a calcareous sand, the whole consolidated as a compact and nearly homogeneous mass; while as we recede from it, the more finely comminuted materials are spread out over the adjacent sea-bottom, and, mingling with a little argillaceous matter, they form the thinbedded argillaceous dolomite of this region."

Hall erred, however, in calling the enclosing strata Waukesha dolomite,[1] for it has since been shown that this bioherm lies in the Racine dolomite.

Chamberlin (1877, p. 360-380; 1883, p. 183-186), in his monumental *Geology of Wisconsin*, was the first investigator who adequately described the Wisconsin bioherms. He pointed out their unique stratigraphic rela-

[1] The Waukesha dolomite, which is not shown on the general stratigraphic column of eastern Wisconsin that appears in this paper, lies immediately beneath the Racine dolomite in the vicinity of Waukesha, Wisconsin. Its correlative relations with the lower part of the Niagaran sequence have not yet been definitely established.

tions, described the different lithological facies associated with them, and listed many of the fossils that occur in the bioherms and enclosing strata; he did not, however, differentiate between the bioherm faunas and those found in the interbiohermal lagoons. Whitfield (1882, p. 267-319) described and illustrated many of the species listed by Chamberlin.

It was largely due to Chamberlin's descriptions that the Wisconsin bioherms became so widely known. He and Salisbury (1906, p. 381-383) used them as a basis for their discussion of Niagaran coral reefs in Volume II of *Geology*, and nearly every subsequent writer on American Silurian coral reefs has referred to Chamberlin's original discussion.

Grabau (1903, p. 341-343) visited a number of Wisconsin bioherms in his investigation of Paleozoic coral reefs, and in his report he paid some attention to reefs at Wauwatosa, Milwaukee, Cedarburg, and Grafton. In later publications (Grabau, 1921, p. 335; 1924, p. 418-420) his brief references apparently are based on observations made during the visit mentioned.

Alden (1906, p. 2) briefly described the several bioherms in the Milwaukee quadrangle and later (Alden, 1918, p. 90-92), in discussing the Quaternary geology of southeastern Wisconsin, called attention to others in surrounding areas. He correlated the bioherms and enclosing strata with the Racine dolomite.

Revival of interest in the Silurian rocks of the Michigan Basin in 1927 was responsible for a number of papers on coral reefs which appeared during the following decade. Cumings and Shrock (1927, p. 9; 1928a, p. 599; 1928b, p. 142) referred to the Wisconsin reefs, comparing them with the Indiana bioherms which they were describing; Fenton (1931, p. 203 ff.) mentioned them in discussing the importance of stromatoporoids as reef-builders in the Chicago region; and brief references appeared in reports or textbooks by Twenhofel (1932, p. 747), Shrock (1929, p. 25; 1935, p. 107-108), Shimer (1933, p. 180), Twenhofel and Shrock (1935, p. 126), and Schuchert and Dunbar (1933, p. 185; 1937, p. 64). It is possible that there are a few other scattered references to the reefs of eastern Wisconsin, but these are the most important.

GENERAL STRATIGRAPHIC COLUMN FOR EASTERN WISCONSIN

The Silurian sequence of eastern Wisconsin, in the region where bioherms are known to occur, is shown in Table 1 (Ehlers, 1921, p. 129-130; Shrock, 1935, p. 107-108).

CHARACTERISTICS OF BIOHERMS

Structures considered to be partly or entirely of biohermal origin occur along the contact between the Mayville and Byron dolomites, at

TABLE 1.—*Silurian formations and contiguous strata in eastern Wisconsin*

DEVONIAN				Fossiliferous, argillaceous limestones unconformably overlying the Silurian at and north of Milwaukee
Unconformity				
SILURIAN	Cayugan		Waubakee dolomite	Gray, thin-laminated, unfossiliferous dolomite unconformably overlying the Racine formation at several localities north of Milwaukee
Unconformity				
	Niagaran		Racine formation	Gray, dense to granular, massive to thin-bedded, barren to highly fossiliferous dolomite containing bioherms in the vicinity of Milwaukee
		Manistique formation	Cordell dolomite (=Upper Coral)	Gray, fairly thin-bedded, cherty, coralline dolomite found mainly north of the latitude of Mayville. The coral bioherms at Quarry and Grimms are probably at this horizon
			Schoolcraft dolomite (=Lower Coral)	Gray, thin-bedded dolomite containing two prominent beds of *Pentamerus*
		Burnt Bluff formation	Hendricks dolomite (=Transitional beds)	Transition series of thin-bedded, semilithographic dolomites and fairly thick bedded, unfossiliferous, granular dolomites
			Byron dolomite	Gray, thin-bedded, mud-cracked and ripple-marked, sparingly fossiliferous, semilithographic dolomite containing small bioherms at several localities
	Medinan (Alexandrian)		Mayville dolomite	Gray, thick-bedded, uneven-textured, cherty, fossiliferous dolomite
Unconformity				
ORDOVICIAN				The Neda oölitic iron ore occurs locally as lenticular bodies between the Mayville dolomite and the underlying Maquoketa shale. Both Neda and Maquoketa are of Ordovician age

several horizons within the Byron dolomite, in the Cordell dolomite, and at several horizons in the Racine formation. These are alike in being structureless masses of uneven-textured, porous dolomite which interrupt the uniform bedding of the regional stratigraphic sequence. They differ considerably, however, in many details of structure, lithology, and paleontology, so they are discussed by formations.

BIOHERMS ASSOCIATED WITH MAYVILLE-BYRON CONTACT

GENERAL CHARACTER

There is a marked lithological difference between the Mayville and Byron dolomites, yet rarely is there a definite surface of separation between them. There is usually a gradual lithological change, through a few inches to a foot or so, from porous, thick-bedded, fossiliferous dolomite upward into dense, thin-bedded, barren rock. At some localities, however, this transition is interrupted by small, lenticular or mound-like swellings of spongy and granular dolomite full of various kinds of fossils. These reefy masses rise from the top of the upper *Virgiana* biostrome of the Mayville or, if the biostrome is not developed, from the upper thick beds and merge gradually with the lowest thin strata of the Byron which arch over them in gentle domes to give individual bedding surfaces a billowy structure. Laterally they grade into typical Mayville in the lower part and thin-bedded Byron in the upper part. They appear, therefore, to bridge the contact between the two formations and they may well represent continuous deposition from the former into the latter.

The reefy masses range from less than a foot to as much as 23 feet thick and average a few tens of feet across, though in some instances they appear to cover several thousand square feet. The thin-bedded strata on their flanks usually show gentle dips (5-15 degrees), but in some of the structures south of Fond du Lac dips greater than 35 degrees occur. These steeper inclinations probably represent steepening of the initial dip by slumping, for on the flanks of some of the mounds the steeply inclined beds are folded into a series of gentle undulations. In the same localities lenticular masses of highly contorted and brecciated dolomite lie in the midst of the flanking strata. These disturbed lenses appear to represent soft materials which slid down the flanks of the mounds.

Not only is the exact age of some of the mounds uncertain, but the origin of many of them also presents a problem. Some may be local prominences built above the sea bottom by algae, though there are no structures now preserved in the reefy rock to support this proposal. Some appear to be shoals of broken shells drifted together in shallow water by currents and waves. These contain considerable granular dolomite, sometimes cross-laminated, which once may have been a clastic calcareous

sand. Still others may represent local prominences on the bottom built up in part by the numerous organisms living there and in part from the materials precipitated chemically out of the surrounding water or washed in as a clastic sand from the neighboring bottoms. The abundance of fossils in some of the mounds and the paucity of them in others make it difficult to evaluate the importance of organisms as builders, but in a number of instances their contributions were large.

MAYVILLE

Small mounds similar in general respects to those just characterized occur along the Mayville-Byron boundary in the large quarry of the Mayville White Lime Works in the SW¼ sec. 1, T. 11 N., R. 16 E., along the Niagara Escarpment about 3½ miles south of Mayville. (Fig. 1).

The thin-bedded Byron has been almost completely removed from a considerable area in the eastern part of the quarry, and the floor shows numerous gentle domes with 2 or 3 feet of relief and circular in outline with diameters of 15 to 20 feet. Cross sections reveal cores of uneven-textured, granular dolomite full of brachiopods (*Eospirifer*) and ostracods (*Leperditia*). These reefy cores, over which thin beds of semi-lithographic dolomite arch in gentle undulations, grade laterally into the basal strata of the Byron and downward into the porous *Virgiana* biostrome which here forms the top bed of the Mayville dolomite. They may represent continuous deposition from late Mayville into early Byron time, but the brachiopods and ostracods show Byron affinities, and the total absence of specimens of *Virgiana,* which is represented by myriads of individuals in the immediately underlying biostrome, further supports reference to the Byron.

OAKFIELD

Several shoal-like mounds of highly fossiliferous material occur along the Mayville-Byron boundary in an abandoned dolomite quarry near Oakfield (Fond du Lac County), in the NW¼ sec. 23, T. 14 N., R. 16 E. The deposit has the form of an irregular lens with a maximum thickness of 6 feet and an areal extent of perhaps 1000 square feet. Over that area, however, it is broken into several individual mounds. It lies upon the upper beds of the Mayville dolomite, which is relatively unfossiliferous, and beneath typical thin-bedded Byron. Its peripheral relations are not well enough exposed to indicate its exact age.

The material of the mounds has the aspect of a coquina with a matrix of finely comminuted shell matter and granular dolomite, both probably of clastic origin, in which are embedded excellently preserved calcified and silicified corals, bryozoans, and brachiopods. The rock looks like

pink marble, is quite soft and porous, and contains a few small dolomite pebbles. There is no apparent stratification. Weathering causes disintegration of the rock with consequent release of the fossils, and in some places the small, well-preserved specimens may be scooped up by the handfuls. All the fossils are small for the species they represent, and study will be necessary to determine whether they are young individuals or diminutive forms. The entire deposit appears to have been drifted together in shallow water by waves and currents which heaped fossils and clastic sands alike into low ridges and mounds.

HAMILTON

One of the most interesting regions of Silurian sedimentation in eastern Wisconsin lies in sec. 10, T. 14 N., R. 17 E., about 4 miles south of Fond du Lac, where U. S. Highway No. 41 ascends the double-stepped, northward-facing Niagara cuesta. Here, over an area of about 300 acres, the thin-bedded Byron strata are thrown into numerous gentle swells and depressions as they rise over concealed mounds and descend into shallow basins between. Some of the swells show relief of only a few feet, whereas others, judging from the nature and inclinations of the domed beds, must lie above mounds that may be tens of feet high. In areal extent they range from less than 10 feet across to a large ridgelike mass which may be as much as a quarter of a mile in total length.

The material of the mounds, so far as can be observed, is massive, gray to pink, uneven-textured, porous or druzy dolomite normally unfossiliferous. Several mounds are exposed in the low ground north of the quarries on either side of U. S. Highway No. 41, and along the lower bench of the cuesta in the NW¼ NW¼ sec. 10. At the latter locality, and in the quarry of the Western Lime and Cement Company a few hundred yards southeast, the upper 15 to 20 feet of the Mayville dolomite exhibits unusual variability in lithology, apparently the result of deposition under widely differing bottom conditions. At one place in the face of the escarpment dark-gray to pink, nodular, slightly argillaceous dolomite changes downward into thick-bedded, uneven-textured dolomite full of cindery fossiliferous nodules of chert in the upper few feet and of poorly preserved stromatoporoid heads in the lower part of the 5-foot unit. Laterally both of these lithologic types grade into a massive 6-foot bed of porous, uneven-textured reefy rock. Elsewhere this reefy horizon is represented by argillaceous, conglomeratic dolomite or gray to buff, coarsely granular dolomite without fossils. Rock of the last type is found in a number of 5- to 10-foot mounds between the escarpment and the large dolomite quarry a short distance to the southeast, and it appears again with considerable development in a small secondary

quarry just south of the lime kilns and southwest of the main quarry. This same type of rock may constitute the ridgelike mound that appears to extend from the small quarry just mentioned eastward nearly to U. S. Highway No. 41. The existence of such a ridge is suggested by two lines of evidence. First, the large east-west trending hill in the center of sec. 10, in the flanks of which several quarries have been opened, appears to be a reflection of such a buried mound, and second, strata on the northern side of the hill consistently dip to the northward whereas those on the southern side dip southward. Unfortunately, the central part of the hill is covered by glacial drift except in a few small exposures where typical thin-bedded Byron appears.

The Byron strata which grade into and override the postulated mound are excellently exhibited in the large dolomite quarry of the Western Lime and Cement Company near the center of the NW¼ sec. 10. The strata in the east and south walls of the quarry dip northward at angles of 5 to 24 degrees and in one small structure where local slumping took place the initial dip has been steepened to over 35 degrees. In an east-west direction the strata show prominent undulations as they dome and sag, apparently in conformity with small buried mounds similar to those a short distance west along the escarpment. At several places in the quarry wall individual beds double or even treble in thickness as they rise toward or grade into the postulated mound. In some of these beds there are numerous undulations and corrugations, produced apparently by incipient sliding of the material while it was still soft. One such exposure in the southeastern corner of the quarry shows several beds, with northerly dip of about 22 degrees, in which intraformational corrugation is excellently developed. Elsewhere in the quarry walls individual beds suddenly swell to twice or thrice normal thickness and change from thin-bedded, semilithographic dolomite to highly deformed and brecciated rock, and along the east wall such a brecciated mass over 20 feet thick and probably twice that distance in diameter interrupts the uniform northward-dipping (7 to 10 degrees) thin-bedded Byron dolomites. These lenses and irregular masses of brecciated rock are believed to have been formed by subaqueous gliding or sliding of soft calcareous muds. The same phenomenon on a much smaller scale also occurs on the flank of a small mound several hundred yards north of the Standard Lime and Stone Company quarry in the NW¼ NE¼ sec. 10. The slopes on which the postulated gliding took place appear to have been inclined from 7 to 12 degrees.

The age and origin of the mounds merit brief discussion. It appears reasonably certain that the mounds were initiated in late Mayville time, and stratigraphic relations with the enclosing thin-bedded dolomites

strongly suggest that their growth was not completed until a considerable thickness of Byron dolomite had been deposited around them and finally over them. Their age, therefore, seems to be in part Mayville and in part Byron. Their origin is less certain. The thin-bedded strata in the shallow basins between the mounds contain a few favosite corals, numerous pentameroid brachiopods, and a great abundance of small ostracods. The massive, unstratified reefy rock of the mounds, on the other hand, is nearly always unfossiliferous and is thoroughly dolomitized. If fossils ever were present, it seems highly probable that they were destroyed during the alteration of the rock. It may be postulated that algae, which could easily have flourished in the shallow waters around the mounds and which do not always leave structural evidence of their contributions, and stromatoporoids, of which a number of poorly preserved heads occur in several places, may have helped to build parts of the mounds. It seems more likely, however, that above and around the mounds chemical conditions favored rapid precipitation of calcium carbonate, and that waves and currents were active in grinding up organic materials abandoned on the bottoms and heaping up the resulting calcareous sands. All evidence strongly suggests that during formation the mounds stood above the general sea bottom as low eminences, and the shallowing of the water over them may have produced conditions of temperature and agitation that favored precipitation, whereas in the immediately surrounding waters, which were a few feet deeper, these same conditions did not prevail.

PEEBLES

A situation similar to those at Oakfield and south of Fond du Lac occurs in a small abandoned quarry along the Niagara Escarpment about a third of a mile south of Peebles, in the W½ sec. 32, T. 16 N., R. 18 E. The entire quarry face is in the Mayville dolomite, but the thin beds that cap the escarpment are apparently Byron. The Pleistocene glaciers ripped off the latter beds from part of the escarpment so that the area directly above the quarry is very irregular with smoothed and striated mounds and shallow basins, all excavated in the reefy bed at the top of the Mayville.

This uppermost unit of the Mayville exhibits variable lithology and thickness. Normally it is a 6-foot bed of dense, bluish-gray dolomite which shows some stratification, but at and near the quarry it swells into a massive, lenticular body of porous and coarsely granular, reefy, fossiliferous dolomite. This lens grades downward into the underlying bedded dolomite. Its base is flat, but the top is quite uneven, and the overlying thin beds of Byron dolomite show marked undulations as they conform with the

irregularities of the surface they cover. Along the top of the escarpment above the quarry the reefy mass thickens from slightly over 6 feet at the escarpment edge to over 23 feet at a point about 200 feet eastward; a similar thickening occurs in a southerly direction in another exposure nearby. In both instances the thin-bedded Byron strata lie upon the surface of the reefy mass and nowhere do they seem to grade into it. These relations suggest that the mound antedated deposition of the Byron.

The lenticular body appears more like a great shoal of fragmented fossils and clastic calcareous sand than a bioherm built *in situ* by organisms. The cross-lamination and general stratigraphic relations of the mass strongly suggest that the material was washed to its present site by currents and waves and later buried beneath the Byron muds. Some of the preserved fossils, however, may well be the remains of organisms which actually lived on the mounds as they were accumulating.

SUMMARY

Bottom conditions during late Mayville and early Byron time appear to have been quite variable from place to place, but all evidence indicates shallow water and fairly vigorous wave and current action. It appears that organisms flourished over parts of the bottoms, where in some instances they helped to build low mounds, and were absent or rare over other parts. Waves and currents apparently destroyed parts or all of some faunal assemblages, grinding up the organic materials and ultimately heaping them together, along with clastic calcareous sand, into low mounds and ridges. Some of these ridges persisted into Byron time and were buried beneath the thin-bedded dolomites that were deposited in shallow waters. Judging from their stratigraphic relations, it appears quite likely that a few of the mounds actually represent continuous deposition from late Mayville into early Byron time.

BIOHERMS IN BYRON DOLOMITE

GENERAL CHARACTER

Numerous small reefy mounds and lumpy masses occur at several horizons in the Byron dolomite. These are cuboidal or hemispheroidal, with flat bases, steep and serrated peripheries, and domed or undulating tops. They are usually less than 10 feet high and rarely over 15 feet in greatest diameter. The rock composing them is either dense, fine-grained dolomite with little evidence of bedding, or a structureless mass of granular or uneven-textured dolomite. Fossils are always rare. Both types of rock contain numerous large pockets of shale and irregular cavities lined with calcite crystals or partly filled with powdery limonite; they weather easily to a porous or cavernous condition. Since the mounds are so much more

susceptible to weathering than the enclosing rock, their positions in the quarry walls are commonly marked by large and irregular holes partly filled with rotten, granular, or cavernous rock.

The mounds are rather sharply separated from the underlying strata along a well-defined base, though in some instances there is a transitional zone (Figs. 2A, 2C, and 3A). Peripherally they give way to thin-bedded dolomites with abrupt or gradual change in lithology. Wedge-shaped masses of reefy rock may extend outward into the enclosing bedded dolomites, and similar extensions of bedded rock may deeply penetrate the mounds (Figs. 3A-C). Against the peripheries of the mounds the surrounding strata rise sharply, with inclinations that in some instances are astonishingly high (30 to 70 degrees) for strata that have not been subjected to severe folding or faulting (Figs. 2B-C). Inclinations of less than 25 degrees are probably initial, but it is almost certain that those over 25 degrees have been steepened by slumping of the bedded calcareous sediments while they were still relatively soft.

The domed or undulatory tops of the mounds are usually rather sharply separated from the thin strata that cover them. These strata, rising gently from the surrounding area, at first undulate in conformity with the upper surface of the mound but within a few feet vertically they become essentially horizontal (Figs. 2A-C).

In one locality the reefy mounds grade laterally into a massive, cavernous bed of uneven-textured dolomite about half the thickness of the mounds. This bed has essentially the same lithology as the reef rock, indicating that during its formation bottom conditions and sediments were similar over considerable areas.

KNOWLES

In two quarries about 1½ miles north of Knowles, in northeastern Dodge County (Fig. 1), numerous small mounds similar to those just characterized interrupt the uniform bedding of the Byron dolomite. These mounds are confined to a 10-foot horizon approximately 35 feet above the base of the Byron, which, incidentally, may be seen in contact with the Mayville dolomite in the Nast Brothers' quarry about three-quarters of a mile south of Knowles, in the SE¼ sec. 30, T. 13 N., R. 17 E.

Twenty-five mounds occur in a distance of a quarter of a mile along the quarry face at the plant of the Western Lime and Cement Company, 1½ miles south of Knowles in sec. 18, T. 13 N., R. 17 E., and an additional dozen appear in the quarry of the Standard Lime and Stone Company less than a quarter of a mile north; all are at the same general horizon and exhibit the same stratigraphic relations.

The foundation layer of the mounds is a massive, cavernous, 2- to 3-

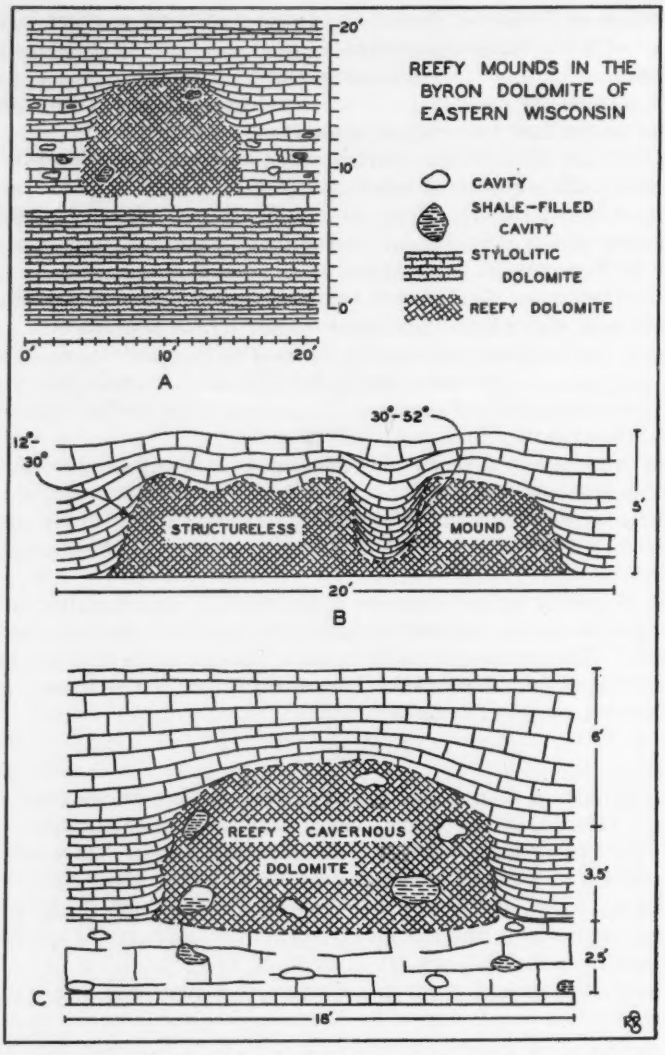


FIGURE 2.—*Biohermal mounds in the Byron dolomite near Knowles*

A. Small cuboidal mass of structureless dolomite in southwest wall of Standard Lime and Stone Co. quarry, 1½ miles north of Knowles (Dodge County), in sec. 18, T. 13 N., R. 17 E.

B, C. Small reefy mounds and associated rocks in west wall of Western Lime and Cement Co. quarry, 1½ miles north of Knowles (Dodge County), in the SE¼ sec. 18, T. 13, N., R. 17 E.

foot transitional bed, which is fine grained and dense at the base where it gives way downward to similar rock that is thin-bedded and stylolitic, and cavernous and granular at the top where it merges with the bases of the mounds or the cavernous bed which extends laterally from the mounds (Figs. 2A, C).

The strata that surround the reefy masses differ somewhat in lithology. The lower part of the sequence comprises a few fairly thick beds of buff, uneven-textured dolomite containing numerous cavities filled with gray shale or powdery limonite or lined with calcite crystals. The beds of this unit may pass beneath or directly into the mounds, but if they show the latter relation they do not rise against the mounds as do the overlying strata (Figs. 2A, C). The upper part of the sequence consists of fairly thin-bedded, even-textured, fine-grained or granular pink dolomites. These strata always rise against the mounds before grading into the reefy mass and contain fewer cavities than the underlying beds.

The beds overlying the mounds are thin and, except for uneven bedding and occasional small cavities, exhibit about the same lithology as the immediately underlying rock.

There are very few fossils in either the mounds or enclosing strata. Ostracods are present in the cavernous bed which grades laterally into the mounds, and a single orthoceratite cephalopod was also found at this horizon. Crudely laminated masses in a few of the mounds may represent organic structures, but their origin is uncertain.

CHILTON

Several reefy mounds are well exposed in a low, northward-facing bluff which crosses County Trunk Highway "G" about a mile south of Chilton (Fig. 1). These lie in the NE¼ sec. 25, T. 18 N., R. 19 E. and the NW¼ sec. 30, T. 18 N., R. 20 E.; they are illustrated in Figures 3A-C. Their stratigraphic position is in the upper 20 feet of the Byron dolomite.

These mounds exhibit crude bedding and a lenticular nature which strongly suggests that they may be elongated ridges. They range from 5 to 15 feet across and probably do not exceed 10 feet in height. They are composed of structureless and cavernous, semisaccharoidal or dense dolomite which merges gradually into the enclosing rock in all directions (Figs. 3A-C). Most of the mounds show interfingering around the margins, and in one (Fig. 3C) the bedded rock deeply penetrates the reefy mass.

The enclosing rock is fairly coarsely crystalline dolomite in alternating thin and thick beds. The thick beds possess extremely uneven texture in some places, thicken and thin considerably within short distances, and in general have a very billowy upper surface. The thinner layers are arched or bent downward to fit this billowy surface so that in an extensive hori-

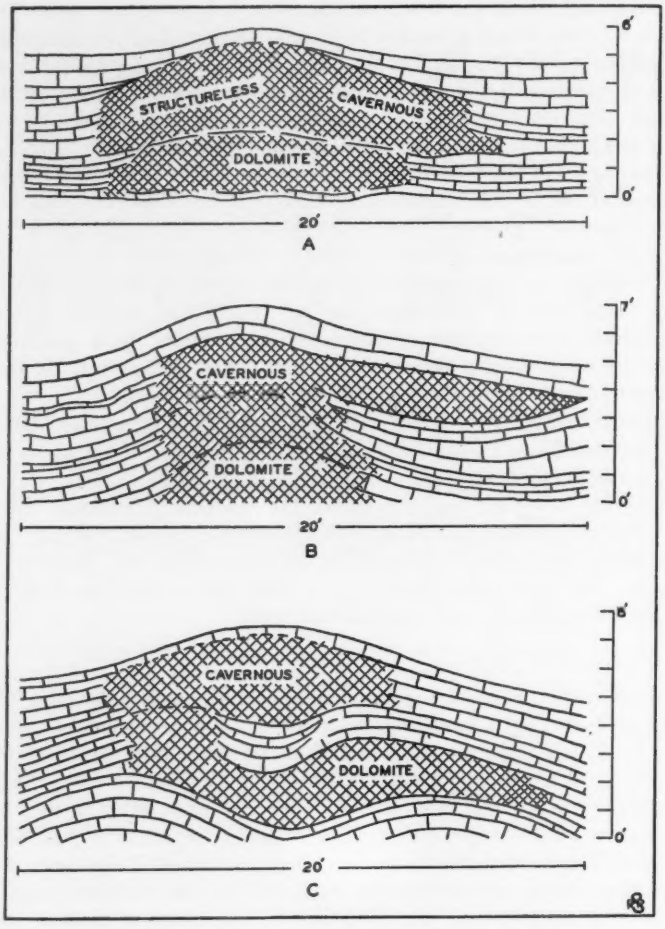


FIGURE 3.—*Small bioherms in Byron dolomite near Chilton*

A-C. Small reefy mounds exposed along a low, northward-facing bluff about a mile south of Chilton, in sec. 25, T. 18 N., R. 19 E. and sec. 30, T. 18 N., R. 20 E.

zontal exposure a bedding surface presents a very billowy appearance. Around some of the reefy swellings the strata are inclined at fairly high angles, in several instances as much as 45 degrees. The beds which dip more than 25 degrees have very likely been steepened by slumping. In a small quarry several hundred yards east of these mounds the thin dolo-

mite layers directly overlying the reefy masses are well exposed in the floor and show some unusual rectangular mudcracks which are associated with the usual polygonal and irregular types.

DOOR COUNTY

At several localities in the Door Peninsula in the upper part of the Byron dolomite a prominent vuggy, uneven-textured, 1-foot bed of dolomite locally swells to twice or thrice its normal thickness and takes on a reefy character. This stratum, sharply separated from the thin-bedded, semi-lithographic, conchoidally fracturing dolomite in the midst of which it lies, is unfossiliferous but contains many small cavities 2 or 3 inches in diameter. The reefy swellings, which are usually about twice as wide as high, are composed of rock similar to the vuggy bed but they present a massive appearance and are generally porous and granular. The enclosing strata merge into the reefy mounds, and the covering beds arch gently over them. Thin layers of laminated shale occur along the boundary between the massive mound and the distinctly bedded enclosing rock, and in one structure such a shale band has a dip of 28 degrees which is apparently initial. No fossils were found in either the vuggy bed or the reefy swellings, but numerous corals occur in the contiguous beds. Mudcracks and cross-lamination in some of the latter strata suggest shallow-water conditions at the time of deposition.

The origin of these mounds, which in some instances are fairly common (a dozen were counted in the wall of a small quarry about 1¼ miles north of the city of Sturgeon Bay) (Fig. 1), is uncertain. They do not appear to be mounds of clastic material drifted together by currents and waves, and the almost total lack of structure makes it impossible to determine if fossils ever were present.

BIOHERMS IN CORDELL AND RACINE DOLOMITES

GENERAL CHARACTER

Introductory statement.—The bioherms of the Cordell and Racine dolomites are essentially similar. They are lithological and faunal complexes of massive and structureless, uneven-textured, fossiliferous dolomite interrupting the relatively uniform bedding of the regional stratigraphic sequence. Once they were islands of life which stood a few tens of feet above surrounding shallow bottoms where life was rather sparse and over which the restless waves and currents shifted calcareous sands derived from the erosion of the reef mounds. At the present time both bioherms and enclosing strata are invariably composed of dolomite which is usually rather pure. It does not appear likely, however, that they were originally built of dolomite, judging from the state of preservation of the fossils.

Lithologically and faunally these bioherms are quite similar to those of approximately the same age in Illinois (Fenton, 1931, p. 203-212), Indiana (Cumings and Shrock, 1927, p. 7-15; 1928a; 1928b, p. 138-164), Iowa, and Ohio (Cumings and Shrock, 1928a, p. 613-614; Cumings, 1930, p. 199-204). Structurally they consist of a massive central mound of uneven-textured, fossiliferous dolomite—the biohermal (reef) core—surrounded by a relatively narrow zone—the biohermal (reef) flank—where thin- or relatively thick-bedded, granular and porous, sparsely fossiliferous strata lap against and grade into the core, often with steep dips. In the peripheral part of the flank the inclined strata become horizontal and grade laterally into the relatively unfossiliferous rock of the interbiohermal lagoons.

No completely preserved bioherms are now fully exposed, but from numerous remnants some conception of their size and structure may be gained. The core and flank together often exceed half a mile in greatest diameter, judging from the exposures at Quarry (Pl. 1C), Grimms (Pl. 1B), and Wauwatosa (Pl. 2). The original thicknesses of the cores cannot be determined, for in no known bioherm are both base and top exposed, but 63 feet of core rock occurs in the Schoonmaker bioherm at Wauwatosa, and it seems reasonably certain that some of the mounds may have been as much as 100 feet thick. These dimensions are of the same general magnitude as those found in the other States referred to,

The fauna which lived on the biohermal mounds was a rich and varied one, whereas that of the flank region was small and restricted. Most of the small number of organisms of the interbiohermal areas were cephalopods and trilobites. Algae seem to have lived both on and around the bioherms.

Biohermal (Reef) core.—The central core of the bioherm is a domal or ridge-shaped mass of cavernous dolomite which is very irregular in structure and texture. As a whole it is often more resistant to erosion than the surrounding bedded rock so that it sometimes is etched out in relief as a low mound. Such hillocks in the Baltic region, and in northern Indiana and Ohio, are designated *Klintar* (Shrock, 1929, p. 17-18).

The main structure of the reef core is a loosely knit, reticulating network of dense, blue-gray dolomite which is hard and tough and which, because of its rigid framework, gives the biohermal mound its strength and its resistance to denudation. In some instances this primary framework is supplemented by lenticular or irregular masses of dolomitized coquina in which cementation of the shell matter has produced a strong and rigid yet highly porous rock. In these latter fossiliferous lenses half or more of the volume may be voids and coated cavities. Wetted surfaces of the biohermal core often show large hemispheroidal or ball-shaped

bodies of concentrically laminated dolomite which may be either thoroughly diagenized stromatoporoid heads or algal colonies. In some of the larger cavities surfaces are frequently coated with explanate masses of thinly laminated dolomite which resemble the flowstone of caves but which are probably of the same origin as the cryptozoan masses of older rocks.

Some parts of the core are composed of dense dolomite that has been brecciated and then recemented; when weathered this rock resembles disintegrated concrete composed of rounded and angular coarse aggregate. The blocks or fragments are often laminated and originally may have been parts of a bedded formation on the biohermal flank or in a depression on the biohermal mound.

The central part of the core rarely exhibits bedding, but in some bioherms the peripheral portion shows crude bedding which radially outward gives way to well-defined stratification in the inclined flanking beds. The inclinations of the latter are usually low, from a few degrees to as much as 25 degrees, and the bedding is frequently uneven.

Joints are irregular, discontinuous, and not very numerous so that weathered exposures of core rock appear unusually massive even though much pitted by numerous cavities.

The cavities vary considerably in size, from a few inches to 3 feet or more, and exhibit marked irregularity of form. They are in some instances partly or nearly completely filled with laminated or massive gray shale. Commonly they are druzy with calcite crystals and often they are coated with banded onyx-like calcite resembling the deposits found on cave walls. In some reefs they are partly filled with granular dolomite, the individual rhombs of which are still sufficiently cemented as to constitute a spongy or highly porous rock, suggesting that the cavity has resulted from the partial removal by solution of the porous granular dolomite that once was an irregular body within the denser surrounding rock. Some of the cavities appear to represent original voids in the biohermal mass, a condition commonly reported on modern coral reefs; others show by their shape and vague impressions that once they were the sites of organic structures (shelled forms predominantly); still others were probably formed by solution as already suggested. These cavities give to the weathered surface of the biohermal core a very rough and cavernous appearance.

Small pockets of quartz sand and laminated shale occur in a few of the cores. These are not unique to the Wisconsin bioherms, however, for Grabau (1924, p. 421) reported similar bodies in the Gotland reefs, and Cumings and Shrock (1928a, p. 601) found them in the Silurian bioherms of northern Indiana. The argillaceous sediment is not unusual for it is common throughout the core as well as in the enclosing strata, but the

sand is an enigma, for there are no sandstones in the Silurian of eastern Wisconsin. The sand may have come from the erosion of Ordovician or Cambrian sandstones in the central part of the State or it may possibly have been drifted in attached to the roots of seaweeds and similar plants, but neither suggestion is very satisfactory.

The core rock is commonly very fossiliferous, and in some bioherms the fossils are so abundant that the rock is essentially a coquina of bryozoan zoaria, coral heads, and molluscan shells cemented into a porous matted mass of considerable rigidity. On one bioherm, in an area of about one square yard, a dozen orthoceratite cephalopods were counted; on another it was impossible to step in any direction without placing a foot on a tabulate coral head. In some bioherms, however, diligent search is required to detect even the faintest vestige of organic structure. In such instances, however, the apparently unfossiliferous rock almost always shows faint organic structures when wetted.

The present irregularity of texture and lithology in the reef core indicates that during its formation and since its burial the core rock has undergone alteration to variable degrees. Except for scattered vugs containing calcite, the core rock is composed of dolomite. Whatever the fossils were composed of in the living state, they now are invariably dolomitized, and as a result the fossil content consists largely of internal and external molds, fillings of these molds, and impressions. The finer and more delicate structural features and surface markings are commonly lacking.

The reef core appears to have been built of material from three sources. Organisms extracted carbonate salts from the sea water and used them in constructing their hard parts, which as complete structures or comminuted fragments became part of the growing bioherm when the organisms died. Wave action around the periphery of the biohermal mound eroded away some of the reefy rock, and the resulting calcareous sand filtered down through the open meshwork of the bioherm to fill some of the voids and also to act as a cement. Finally, the metabolic processes of the organisms and the physical and chemical changes wrought in the waters by temperature and pressure variations must have set up chemical conditions favorable for precipitation of carbonates. These precipitated carbonates aided in cementing the cavernous reef mass into a stronger and more compact mound and also contributed to the accumulations on the biohermal flank and in the interbiohermal lagoons. Present stratigraphic relations indicate that the bioherms were probably low mounds which stood a few tens of feet above the surrounding bottoms and which were densely populated areas where wave and current action was vigorous.

The bases of the bioherms, where they have been seen (Pl. 2), are clearly distinguishable from the underlying bedded dolomites. In the

Schoonmaker bioherm there is a slight downward flexure under a part of the core, as though the underlying strata had been bent downward somewhat from the weight of the overlying rock. This same relation has been reported from reefs elsewhere, but whether the explanation just given is the correct one for the flexure is not definitely determinable.

Laterally the core rock may change abruptly into the inclined, well-bedded rock of the biohermal flank, or there may be a gradual transition between the two, so that it is difficult to determine where one type of lithology ends and the other begins. In Figure 4 and Plates 1 and 2 this change is shown as an abrupt one, because of the symbols used, but it should be emphasized that in most instances the change is transitional.

Biohermal (Reef) flank.—The biohermal flank constitutes a relatively narrow zone concentric to the core in which well-bedded, granular, and porous dolomites dip away from the core at angles of a few degrees up to as much as 54 degrees. These inclined beds pass into the core on the one side and outward into the strata of the interbiohermal lagoons on the other. In some instances they interfinger with the core rock and often they thicken conspicuously as they approach the core. Where they rise to surmount the biohermal mound they show gentle doming and soon gain a horizontal position above the core (Pl. 1A, B). It is assumed that dips in excess of 25 degrees represent steepening by slumping, compaction, and similar movements. Such certainly appears to be the case in parts of the Schoonmaker bioherm (Pl. 2).

The material of the flank rock is normally massive but well-bedded, evenly textured, porous or cavernous, granular dolomite with few fossils. Often it has the appearance of a poorly cemented calcareous sand. Fossils are present in some instances, as at Cedarburg (Fig. 4), and are usually represented by cavities which mark where crinoids and coral heads were once buried. As a rule, however, the flank rock is unfossiliferous, and this should be the case if the material in it came, as it does in modern reefs, from the grinding up of biohermal core material.

Outward the flank strata grade into the interbiohermal strata, often with transition and always with considerable lithological change. In a few bioherms cherty nodules appear at the periphery of the flank area, suggesting that there is where the quieter waters of the interbiohermal lagoons began, for it does not seem very likely that the colloidal silica of which the cherty nodules are composed could have been precipitated over the biohermal mound where the water was probably always more or less agitated (Pl. 1C).

Interbiohermal (Interreef) lagoonal deposits.—The normal Cordell sequence in the area where bioherms have been found consists largely of

thin- and irregular-bedded, fairly dense, coralline dolomites with scattered chert nodules. In the biohermal areas the normal strata tend to be less fossiliferous than they are where bioherms are absent, and this may suggest that there were deeper waters in the latter places or at least different bottom conditions. Furthermore, the chert nodules which occur in the Cordell beds immediately surrounding bioherms appear to be of syngenetic origin, whereas the chert, much of which occurs as irregular masses, stringers, and layers, that is found in the Cordell far away from the biohermal areas where biostromal layers are common, appears to be in large part of epigenetic origin. The epigenetic origin for the latter chert is also indicated by the fact that shells of *Pentamerus oblongus* and coralla of tabulate corals are usually silicified.

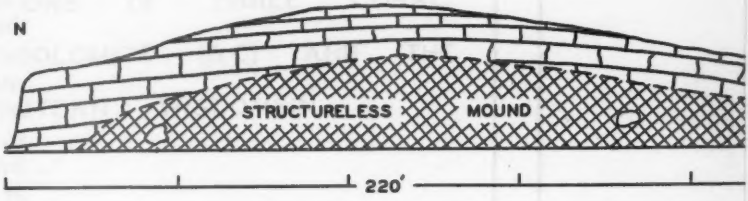
The interbiohermal strata of the Racine formation are either thin-bedded, dense, somewhat argillaceous and sparingly fossiliferous dolomites, as in the Milwaukee area (Pl. 2), or thick-bedded, granular and porous, highly fossiliferous dolomites, as in the Grafton and Cedarburg localities (Fig. 4 and Pl. 1). It is almost certain that the bioherms of the former area lie at a somewhat lower stratigraphic level than those of the latter, and this difference of horizon may account for the different lithologies of the interbiohermal deposits. It is interesting to note, however, that the core rock is essentially the same in the two areas.

At Racine, where biostromal layers alternate with barren beds, and in Door County, where the marble-like Racine is only sparingly fossiliferous, there does not seem to have been biohermal development; it is of further interest that no important bioherms are known in either the Byron or Cordell strata in the Door County area. From these regional relations it appears that biohermal bottom conditions prevailed throughout much of Silurian time over the area between the latitudes of Racine and Green Bay, but that north and south of this area bottom conditions favored biostromal development or were unfavorable for most forms of life. It should be pointed out also that the known occurrences of the Neda oölitic iron ore lenses, which lie between the Mayville dolomite and the Maquoketa shale (Table 1), are essentially limited to the biohermal area (Thwaites, 1914, p. 541). Perhaps shallow-water conditions already existed in this area in late Ordovician time.

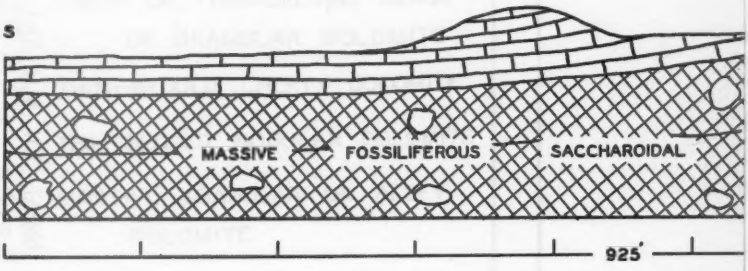
BIOHERMS IN CORDELL DOLOMITE

QUARRY

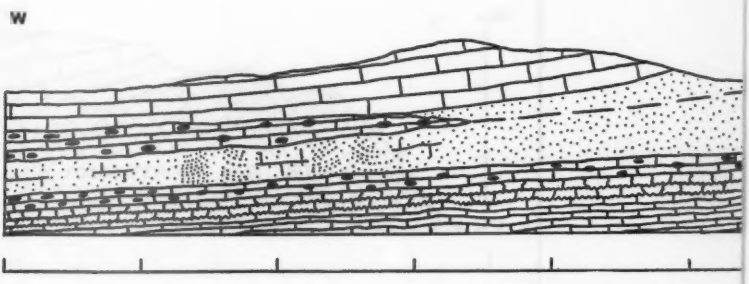
One of the most completely preserved coral bioherms in eastern Wisconsin is unusually well exhibited in two large quarries that have been opened in a prominent hill (klint) just west of the small village of Quarry, Manitowoc County, in sec. 36, T. 19 N., R. 21 E. (Fig. 1). The hill owes



A. MOUND IN ABANDONED QUARRY ON EAST SIDE OF RIVER AT END OF GRAFTON, OZAUKEE COUNTY

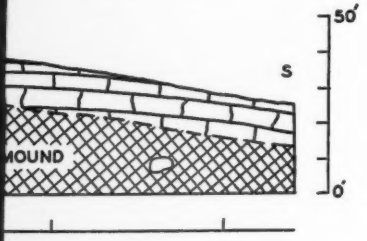


B. PART OF A CORAL BIOHERM IN LIME QUARRY AT GRIMMS,



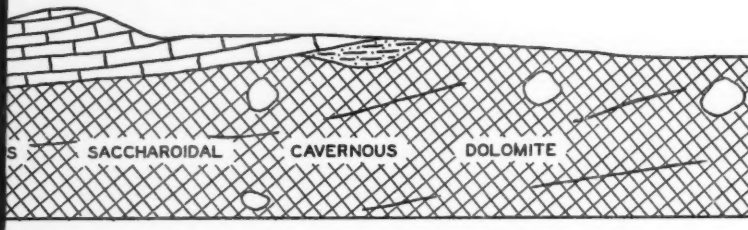
C. BIOHERMAL (REEF) COMPLEX IN THE LIME QUARRIES AT QUA

DIAGRAMMATIC GEC



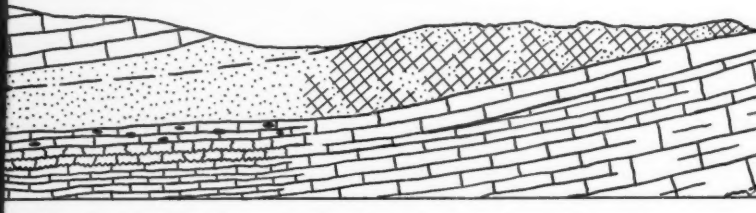
DIAGRAMMATIC
BIOHERMS IN
RACINE DOLOMIT

ST SIDE OF RIVER AT SOUTH
UKEE COUNTY



925'

QUARRY AT GRIMMS, MANITOWOC COUNTY, SHOWING MASSIVE



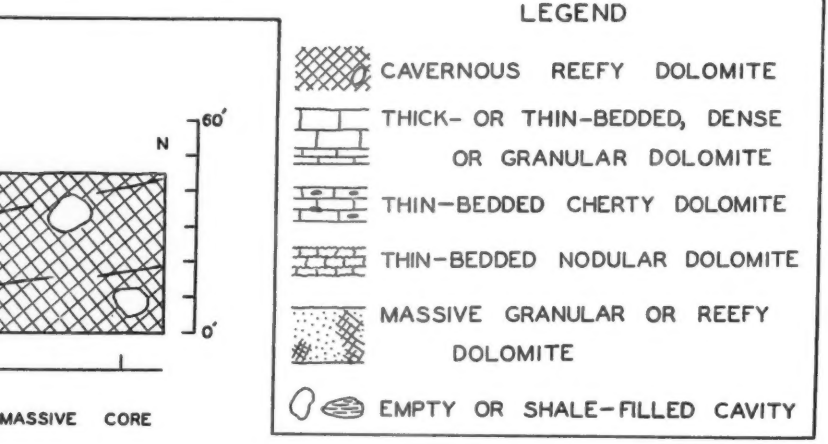
1400'

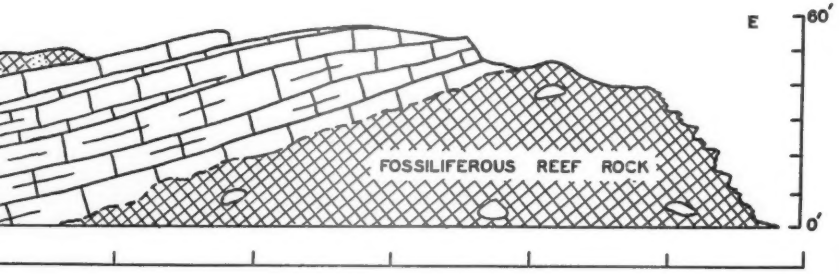
LIME QUARRIES AT QUARRY, MANITOWOC COUNTY, SHOWING LITH
INTER-REEF ROCK

DIAGRAMMATIC GEOLOGIC SECTIONS OF CORAL BIOHERMS OF E

TIC GEOLOGIC SECTIONS OF THREE CORAL
IN THE CORDELL DOLOMITE (B,C) AND THE
OLOMITE (A) OF EASTERN WISCONSIN

1938





NG LITHOLOGIC VARIATION FROM REEF CORE ACROSS FLANK TO

MS OF EASTERN WISCONSIN

its preservation more to the resistance of the thick-bedded flanking strata than to the core rock. Along the quarry walls the beds of the regional stratigraphic sequence may be traced into the rock of the reef flank, and the latter into the massive core. Such a section, based largely on measurements made along the north wall of the northern quarry in 1930, is reproduced in Plate 1C.

The biohermal core, which is only partly preserved as a northeast-trending ridge along the eastern side of the hill, is a structureless mound of rotten, highly porous or cavernous, granular or uneven-textured dolomite full of tabulate corals, stromatoporoids, and shelled organisms. The chief components of the fauna are the first-named organisms, among which the most common are *Favosites, Halysites,* and *Arachnophyllum.* Poorly preserved stromatoporoids are also numerous; many of them measure more than a foot across. In some parts of the mound coral and stromatoporoid heads lie so close together that it is almost impossible to step in any direction without placing a foot on one of them. Many of the heads have been dislodged from the place of growth and are now scattered at random through the core rock, often lying upside down in the granular matrix.

Lapping against the core, with dips as high as 20 degrees, are evenly bedded, rather dense dolomites which have very few fossils. Some of the layers thin toward the core, whereas others thicken, so that in the middle of the biohermal flank the individual beds appear to interwedge. Laterally these flanking strata decrease in total thickness and grade into the more thinly bedded dolomite of the surrounding region. The lowest strata exposed are 16 feet of light-buff, saccharoidal dolomite in fairly thick beds and without many fossils. Overlying this unit is a 4-foot unit of nodular, blue dolomite separated into thin and irregular layers by shale films. Upward this rock gives way to brown, fairly dense dolomite in thin layers with numerous nodules of gray or white chert. The separation plane marking the top of the cherty unit rises gently toward the core, separating the flanking strata already described from the overlying granular and reefy dolomite (Pl. 1C).

This sequence is overlain, away from the bioherm, by a 10-foot massive bed of coralline, porous, and granular dolomite which contains a few druzy mounds of uneven-textured, rotten dolomite similar to that of the biohermal core. Around these small mounds there is faint stratification gently dipping away from them in quaquaversal fashion. As this bed approaches the biohermal flank it changes to saccharoidal dolomite or calcareous sand, and upon the flank it becomes a structureless wedge of reefy dolomite somewhat similar to the rock of the core itself except that it is relatively unfossiliferous. Laterally this wedge thins conspicuously to the

10-foot unit already described and also interwedges with a 7½-foot bed of cherty brown dolomite overlying the 10-foot bed.

The highest strata in the region are thick-bedded, somewhat cavernous, bluish-gray, dense dolomites containing numerous specimens of *Halysites*. These beds, which are present along the western side of the hill, ride up over the reefy wedge and probably represent the flank strata formed in large part from the erosion of a biohermal core, of which only the peripheral part is still preserved.

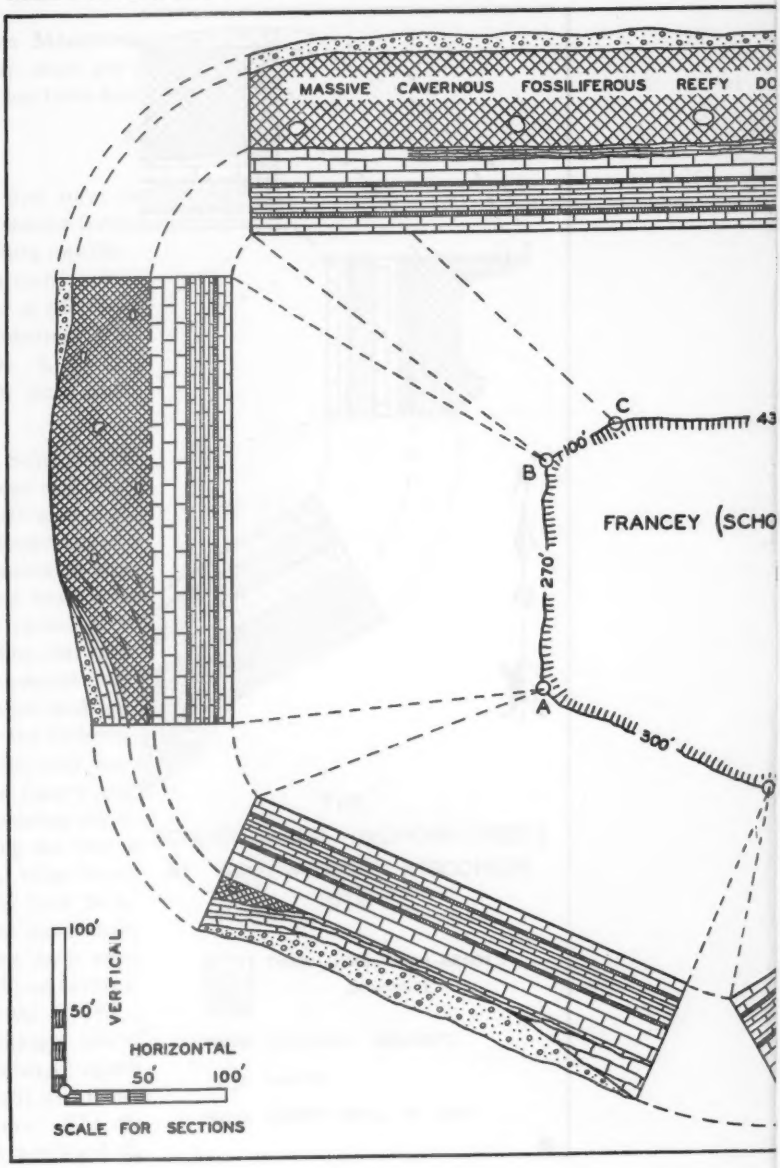
The cherty nodules that occur in the interbiohermal strata appear to be syngenetic and lie within the beds as well as along the separation surfaces. No extensive silicification was found anywhere near the bioherm. It is important to note that the chert leaves off abruptly where the strata of the interbiohermal environment grade into the inclined beds of the flank. This relation suggests that the waters over the biohermal core and flank were too agitated to permit precipitation of colloidal silica, whereas in the quieter waters of the lagoons this material could be deposited.

GRIMMS

A large dolomite quarry has been opened for the production of lime in a small klint just southwest of the village of Grimms, Manitowoc County, in sec. 6, T. 19 N., R. 22 E. (Fig. 1). Along the western face of the quarry is a large coral bioherm covered by thick-bedded dolomite (Pl. 1B). The core rock is gray or light buff, porous, saccharoidal dolomite full of corals and other fossils, most of which are now completely dolomitized. It has a crudely bedded appearance with the beds ranging in thickness from 3 to 15 feet. There are many large calcite-lined druzes and cavities (Pl. 1B). Jointing is poorly developed, and the rock breaks into large wedges or irregular blocks and slabs. The main fossils are tabulate corals of the genera *Favosites* and *Halysites* and stromatoporoids of the hydrozooid type. At the top of the core, in the northwest part of the quarry, there is a large pocket of purplish shale and reddish-brown shaly sandstone. This deposit appears to be material that was swept into a depression or cavity in the biohermal mound from the surrounding bottom, though where the sediment came from is an enigma.

Overlying the core and clearly separated from it are fairly thick beds of gray, granular dolomite much like the core rock except that it is denser and much better bedded. The rock is somewhat fossiliferous and bears essentially the same fauna as that of the core. The beds dip gently off the higher part of the biohermal mound. The hill appears to have existed as a klint before the Pleistocene for the glaciers of that age passed over it leaving the surface striated and polished.

It is known that in the area north of the Quarry-Grimms locality, in


MASSIVE CAVERNOUS FOSSILIFEROUS REEFY DO
C
B
A
100'
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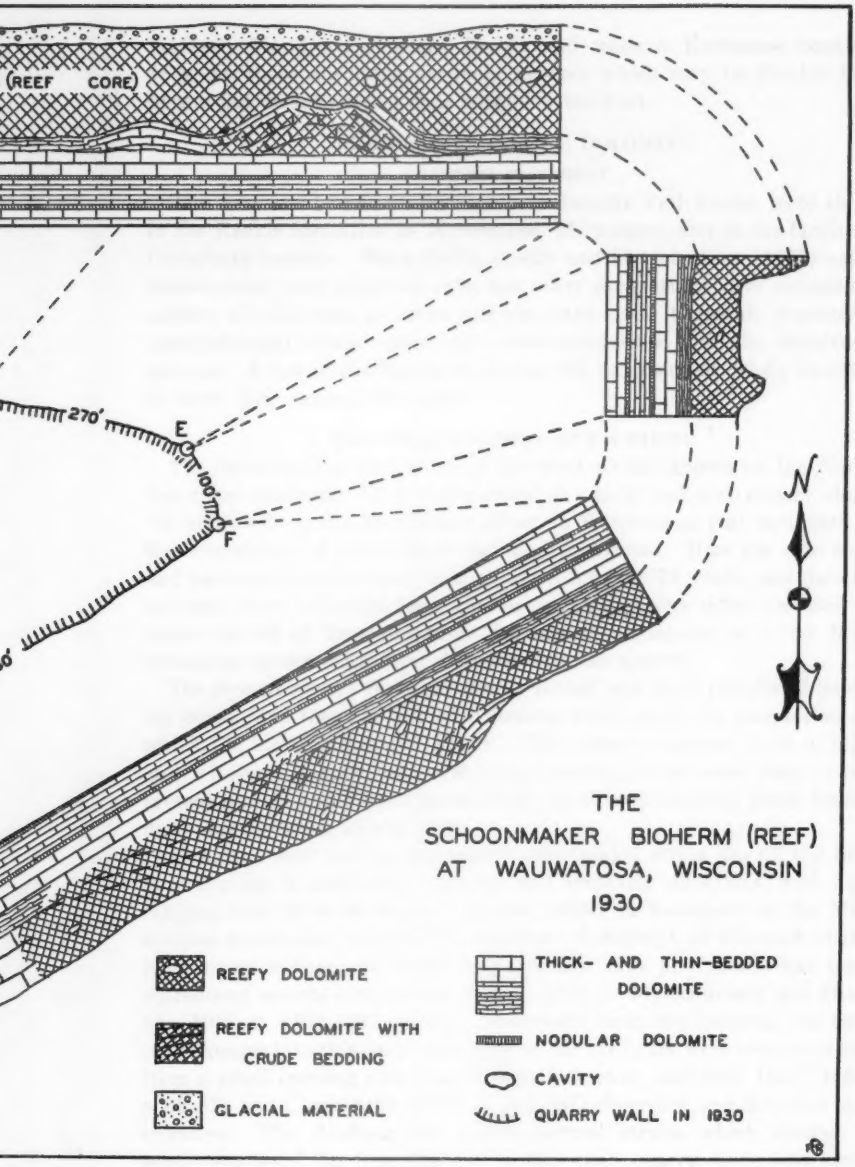
SCALE FOR SECTIONS

SCHOONMAKER BIOHERM


(REEF CORE)
270'
E
100'
F
N
THE
SCHOONMAKER BIOHERM (REEF)
AT WAUWATOSA, WISCONSIN
1930
REEFY DOLOMITE
REEFY DOLOMITE WITH CRUDE BEDDING
GLACIAL MATERIAL
THICK- AND THIN-BEDDED DOLOMITE
NODULAR DOLOMITE
CAVITY
QUARRY WALL IN 1930

HERM

northern Manitowoc, eastern Brown and western Kewaunee counties (Fig. 1), there are numerous reefy mounds which may be Cordell bioherms, but these have not been studied in the field.

BIOHERMS IN RACINE DOLOMITE

GENERAL STATEMENT

The first coral reefs of Wisconsin to become well known were those in the Racine formation at Wauwatosa, Milwaukee, and in the Grafton-Cedarburg locality. Since Hall's (1862) and Chamberlin's (1877) early studies many new quarries, cuts, and other excavations have revealed a number of additional bioherms and also have made it possible to examine interbiohermal strata essentially contemporaneous with the biohermal mounds. A few of the Racine bioherms will be described briefly in order to show their general character.

SCHOONMAKER BIOHERM AT WAUWATOSA

The Schoonmaker reef is one of the most widely known on the North American continent. It is well exposed in a large and deep quarry along the north side of the Menominee River, in Wauwatosa, just northeast of the intersection of State Street and Eighth Avenue. Here the core rock and associated strata have been quarried for over 75 years, and paleontologists have collected from the richly fossiliferous dolomites over a longer period of time, for there are natural exposures in a low bluff extending eastward for some distance from the quarry.

The present quarry faces excellently reveal core rock, peripheral flanking strata, and the thin-bedded dolomites which form the foundation on which the bioherm started (Pl. 2). The bioherm appears to be a high and relatively narrow east-west ridge starting at the west edge of the present quarry and extending eastward for several hundred yards before disappearing under glacial drift.

Along the west wall of the quarry well-bedded strata dip off the biohermal ridge in northerly, westerly, and southerly directions, with dips ranging from 20 to 36 degrees. Alden (1906), in his report on the Milwaukee quadrangle, included an excellent photograph of this part of the bioherm as it appeared about 30 years ago; this photograph has been reproduced several times since (Alden, 1914, p. 90; Schuchert and Dunbar, 1933, p. 185; 1937, p. 64). Westward from the bioherm the core rock changes into thin-bedded dolomites, for the latter were once quarried from a small opening just west of Eighth Avenue, and both Hall (1862, p. 62-63) and Chamberlin (1877, p. 366-367) described and sketched this situation. The flanking and interbiohermal strata, which showed a gentle westward dip from the reef to the small quarry, have long since been quarried away or covered.

The north and south walls of the quarry reveal thick sections of massive core rock resting with apparent conformity on thin-bedded argillaceous dolomites. The latter strata are bent downward somewhat in the south wall as though they had failed slightly under the load of the massive bioherm (Pl. 2, section F-G). More core rock is exposed along the eastern wall, and a low cliff of similar rock extends eastward for several hundred feet in a natural exposure which finally disappears under glacial drift.

The detailed structure of the Schoonmaker bioherm is shown on Plate 2 by a number of cross sections based on measurements made in 1930-1931. Almost continuous excavation during the last 8 years has removed some of the facies that were then observed and has also revealed new aspects of the bioherm; hence the sketch here produced is of composite nature.

The biohermal core is essentially one homogeneous body of uneven-textured, cavernous, highly fossiliferous dolomite. The ridge which it constitutes shows little evidence of bedding or stratification except at the base in the northeast wall (Pl. 2, section D-E), poor and irregular jointing, and crudely serrated margins where the core rock loses its identity in the bedded flanking strata. Once initiated the bioherm apparently grew continuously until its burial or its abandonment by the life that made it; there is no evidence within the core suggesting cessation of growth due to temporary burial.

The greatest thickness of core rock observed was 63 feet, and in this section, which was along the northwest wall, only the base was visible. It is highly probable, judging from the Schoonmaker and neighboring bioherms, that some of the ridges and mounds may have been as much as 100 feet thick. The fossiliferous core rock probably furnished most of the fossils gathered by the early collectors, and it may be assumed that species reported from Wauwatosa came in large part from the core rock of the Schoonmaker bioherm. Since no careful study has been made of existing collections from Wauwatosa, it is not possible to give a complete list of the bioherm-building organisms. Species of the following genera, however, are numerous, and the list will indicate the general character of the fauna:

Corals: *Favosites, Halysites, Heliolites,* and *Syringopora*
Stromatoporoids: *Clathrodictyon*
Bryozoans: *Fenestella*
Brachiopods: *Conchidium, Eospirifer, Rhynchotreta, Schuchertella,* and *Leptaena*
Mollusks: *Amphicoelia, Mytilarca, Streptomytilus; Crateroceras, Cyrtorizoceras, Dawsonoceras, Geisonoceras, Geisonocerina, Hexameroceras, Kionoceras, Leurocycloceras, Monocyrtoceras, Ophidioceras,* and *Phragmoceras*
Trilobites: *Bumastus, Calymene, Cheirurus, Dicranopeltis, Encrinurus,* and *Sphaerexochus*

Laterally the massive core first becomes crudely bedded and granular and then grades into the well-defined flanking strata which dip gently off the biohermal mound, thinning out wedge-like as they give way to the thin-bedded, horizontal strata of the interbiohermal lagoons. Some of these wedges of flank rock are 20-25 feet thick at the junction with the core and thin to less than 5 feet before losing their identity in the interbiohermal beds. In some parts of the bioherm, core and flank rock interfinger and give to it a serrated margin, but in other parts they change from one to the other abruptly, and in these instances the interbiohermal strata end abruptly against the core remaining horizontal rather than rising gently toward it.

These relations indicate that the growing bioherm which stood a few tens of feet above the surrounding bottoms had an irregular margin. In one place there appears to have been a cliff, at the base of which granular calcareous sand worn off the growing bioherm gradually built up into thick wedges of what is now flank rock. In other parts of the bioherm extensions of the core were built out over the adjacent bottom; these were later buried beneath calcareous sand so that now core and flank rock interfinger.

The rock of the biohermal flank is thick-bedded, even-textured, porous and granular, relatively unfossiliferous dolomite which appears to have been formed in large part of the calcareous sands produced from erosion of the growing bioherm.

The base of the bioherm is fairly flat except along the southeastern wall where the foundation strata are bent downward slightly (Pl. 2, section F-G) and along the northeastern wall where the underlying strata are bent upward into several prominent folds (Pl. 2, section D-E). In the latter place the rock has the lithology of the core but bedding like that in the rocks below.

The foundation upon which the bioherm began and the interbiohermal rock which is essentially contemporaneous with the core consists of thin-bedded, even-textured, somewhat argillaceous dolomites which show remarkably smooth separation planes and well-developed joints. The sequence is broken, however, by a few nodular layers, rarely over 1½ feet thick, which may end abruptly and be replaced by irregular pockets of laminated shale or stratified quartz sand. Two such nodular beds are exposed in the deepest part of the quarry, and half a dozen similar ones occur in the Manegold and Story quarries a mile or so to the southeast where thick sections of interbiohermal strata are exposed. These unusual beds are deposits of a disturbed bottom and may indicate storms of abnormal intensity which produced waves sufficiently powerful to rip up the bottom deposits.

The Schoonmaker bioherm apparently stood up as a prominent hill (klint) before the Pleistocene, for the glaciers of that time abraded it deeply before burying it under their deposits. Bouldery till fills in the irregular concavities gouged out by the ice and forms a thin mantle over the biohermal mound. It is overlain in turn by outwash gravels, sands, and silts in which the stratification dips gently westward, apparently in conformity with the slope of the old hillside on which the sediments were deposited.

MILWAUKEE AREA

General statement.—There are a number of Racine bioherms in the vicinity of Milwaukee (Fig. 1). One of these—the Schoonmaker—is well exposed and has already been described in detail. The remainder are not well exposed and merit only brief discussion.

Hales Corners.—Several small quarries have been opened in a biohermal facies of the Racine dolomite in the SW¼ sec. 28, T. 6 N., R. 21 E., in southwestern Milwaukee County (Fig. 1). Lenticular beds of porous and granular dolomite exhibit considerable variation in dip and appear to be lying on small mounds of reefy rock. In several openings these beds actually interfinger with biohermal core rock. Local dips as great as 10 degrees occur in several places, and a number of small biohermal mounds 10 to 15 feet across contain large numbers of coral heads, crinoids, and trilobites.

Moody's quarry.—There were once extensive exposures of fossiliferous biohermal core rock along the north side of the Menominee Valley in southern Milwaukee where 25th Avenue crosses the Rapid Transit lines (Fig. 1), in the SE¼ sec. 25, T. 7 N., R. 21 E. A quarry known as Moody's was operated here at the time Chamberlin (1877, p. 365) was making his investigations and he referred to it briefly, but all traces of it have long since vanished before the advance of building activities, and today only a few small exposures remain. There is a low cut about 20 feet high and 140 feet long on the north side of the Rapid Transit tracks and a few smaller exposures on the south side.

The rock now exposed is massive, uneven-textured, cavernous dolomite full of fossils and with vague bedding that dips eastward at about 35 degrees. The fauna is similar to that of the Schoonmaker bioherm. A thick mantle of glacial drift conceals most of what appears to be a large biohermal mound (Chamberlin, 1877, p. 365), and the visible rock surface, which has been grooved and polished by glacial abrasion, is inclined to the south toward the Menominee Valley.

National Military Asylum.—A partly denuded bioherm is exposed as a prominent northward-facing bluff in the grounds of the National Military Asylum on the southwest side of the Menominee Valley in sec. 26, T. 7 N., R. 21 E. (Fig. 1). The exposure is about 300 feet long and 30 to 40 feet high. The rock is typical core material with uneven texture, numerous cavities, and vague bedding which inclines northeastward toward the valley at about 35 degrees. There is some question, however, as to whether this apparent bedding represents true stratification. Fossils are rare.

This bioherm was first mentioned by Chamberlin (1877, p. 365) and has been referred to subsequently by several investigators.

North Milwaukee.—A biohermal facies of the Racine dolomite is poorly exposed in a small water-filled quarry near the center of sec. 26, T. 8 N., R. 21 E. (Fig. 1). The rock is massive, uneven-textured, cavernous dolomite, of which only a few feet can now be seen around the edge of the quarry. The surface of the exposure is uneven and has been grooved and polished by the Pleistocene glaciers. Alden (1906) showed this quarry, which has been referred to as Zautke's, on the geologic map of the Milwaukee Special Folio.

West Granville.—Biohermal Racine dolomite outcrops in several small mounds about 1⅛ miles east of the village of West Granville in secs. 20 and 29, T. 8 N., R. 21 E. (Fig. 1). The mounds rise from 5 to 8 feet above the surrounding lowland and are composed of uneven-textured, porous and granular, fossiliferous dolomite which weathers into calcareous sand. The biohermal rock is underlain by slightly denser rock in layers that average about 4 inches thick. This small area is indicated on the geologic map of the Milwaukee Special Folio (Alden, 1906).

South Germantown.—A biohermal facies of the Racine is well exposed in several quarries at South Germantown in sec. 22, T. 9 N., R. 20 E. (Fig. 1). Chamberlin (1877, p. 363-365) described this area in considerable detail, and later writers have referred to it without much comment. The massive beds of somewhat cavernous dolomite show variable dips which apparently reflect the biohermal mounds that underlie them or into which they merge laterally. Chamberlin sketched a large bioherm at this locality, but exposures apparently are not so extensive now as when he studied the area over 60 years ago; hence such a structure is no longer visible.

CEDARBURG—GRAFTON AREA

Coral bioherms are well developed in the vicinity of Cedarburg and Grafton, in southern Ozaukee County (Fig. 1), where they constitute a

lithologic facies of the Racine formation somewhat different from that in the Milwaukee and Wauwatosa areas, and apparently stratigraphically higher. The biohermal faunas also contain some species which are absent in the area farther south.

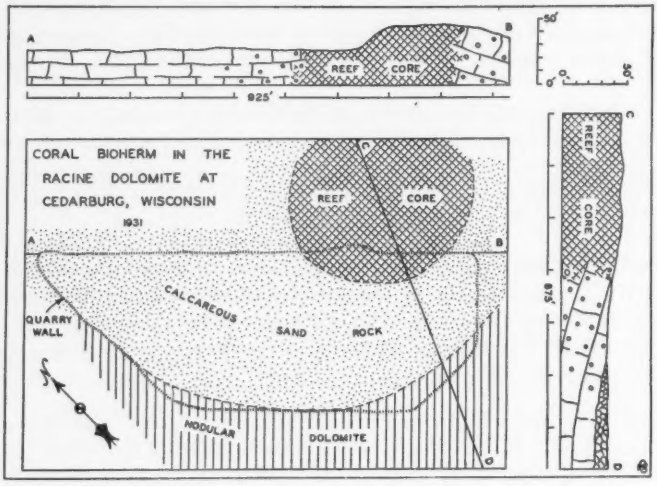


FIGURE 4.—*Coral bioherm in Racine dolomite at Cedarburg*

Probably the best displayed of the numerous bioherms is that in the city of Cedarburg, in the W½ sec. 35, T. 10 N., R. 21 E., where a large quarry has been opened in the thick-bedded rock of the biohermal flank and interbiohermal area. Here a part of the biohermal core is well exposed in the southeastern part of the quarry, and the strata into which the core rock grades are excellently exposed in the walls of the quarry. There is a rather abrupt transition from the biohermal core into the surrounding rock and also a marked faunal change. Lithological relations are shown in Figure 4, which is based on a survey of the quarry and environs as they were in 1931.

The rock of the biohermal core is ragged-appearing, massive dolomite with uneven texture that is due partly to the variation in grain size and partly to the many small and large tubes and holes left after the weathering out of fossils. The rock is blue on fresh surfaces but soon becomes discolored from weathering and breaks along poorly defined joints into large and irregular blocks and slabs with jagged edges. Some parts of the core weather into crumbly, granular material resembling coarsely

textured meal. In one part of the core there is a cavity large enough for a man to enter, and large numbers of corals occur in the wall, suggesting that possibly the cavity represents an original cove in the bioherm that never was filled with sediment.

The core rock is highly fossiliferous with corals, stromatoporoids, bryozoans, brachiopods, and mollusks very much in evidence. There are many heads of *Favosites*, some as much as 8 inches across, *Halysites* of about the same dimensions, and large dendritic growths of *Eridophyllum* and *Syringopora* as large as bushel baskets. In one spot with an area of scarcely a square yard 12 orthoceratite cephalopods were counted. Crinoid stems are everywhere abundant, but strangely enough not a single calyx was found after a diligent search. Columnals are also very abundant in the flanking strata, but there also calices are missing. It is possible that predatory organisms destroyed the calices, or the latter may have abandoned the stem in maturity and floated away.

The biohermal core grades laterally into very pure, even-granular, milky white to buff dolomite in thick beds which dip away from the core at angles of 2 to 5 degrees. The lithologic change is fairly abrupt though hardly as abrupt as the diagram (Fig. 4) would suggest.

The flanking strata contain myriads of small tubes and holes of various shapes. The tubes were formed by the removal of crinoid stems and roots, whereas the holes, of which some are as large as grapefruit, were once occupied by coral heads. There are also some geodized and crystal-lined cavities and a few that are partly filled with granular dolomite. The individual beds, which are less well marked at the periphery of the core than farther out, range from 2 to 5 feet thick and are remarkably homogeneous. The rock apparently is a calcareous sand of unusual purity which was formed by waves and currents grinding up parts of the growing bioherm.

A formation of nodular dolomite of uneven texture and with very irregular bedding overlies the beds just described but is no longer present near the core. It is almost certain, however, that it once covered the entire bioherm. Weathering causes this rock to disintegrate into a concrete-like mass made of bouldery fragments of hard and tough, dense blue dolomite. In unweathered rock there is considerable argillaceous material between these nodular masses. The junction with the underlying granular dolomite is marked and somewhat uneven though probably not unconformable. Since *Halysites* was found in the rock it is assumed to be of Silurian age, but later than the bioherm-building period. In the northwestern corner of the quarry this unit attains a thickness of 30 feet and is responsible for a prominent rocky mound. In every direction therefrom, however, it thins away rapidly because of removal by erosion.

Similar rock occurs in a quarry about 2¼ miles south near the center of sec. 10, T. 9 N., R. 21 E., where it overlies massive, fossiliferous dolomite resembling the flank rock of the northern quarry and apparently representing the interbiohermal sequence of the region. The fauna of the latter dolomite contains numerous species of *Favosites, Halysites, Dinobolus, Pentamerus, Trimerella, Diaphorostoma, Hormotoma,* and *Megalomus,* and has been referred to the Guelph by some investigators.

There are numerous other bioherms in the general vicinity of Cedarburg and especially along the banks of the Milwaukee River between Cedarburg and Grafton, where several quarries reveal good sections of the rocks. One of these, along the east bank of the river, in the NE¼ sec. 25, T. 10 N., R. 21 E., has been quarried for lime, and the present quarry face shows a good section through a small biohermal mound (Pl. 1A). The structureless biohermal rock grades laterally and upward into thick beds of dolomite which are somewhat more even textured. There are very few fossils in this particular bioherm. Numerous other bioherms in the general Cedarburg-Grafton area were well described by Chamberlin (1877, p. 378) and show essentially the same relations and character as those just discussed.

SAUKVILLE

There is a large bioherm in the SW¼ sec. 23 and the N½ sec. 26, T. 11 N., R. 21 E., and several smaller ones in the eastern half of the NE¼ sec. 26 (Fig. 1). Small quarries have been opened in some of the latter bioherms, and the general site is probably that referred to as Noessen's quarry by Chamberlin (1877, p. 378), who called attention to the abundance of crinoids and corals in the rock. The rock in this general vicinity is considerably more calcitic than is usual for the Racine dolomite. It is essentially a crinoid coquina, and the plates and columnals still exhibit calcite cleavage. The matrix of the rock is saccharoidal dolomite.

A small klint of reefy Racine dolomite occurs about 1½ miles southwest of Saukville and is transected by the road which divides sec. 1, T. 10 N., R. 21 E. from sec. 34, T. 11 N., R. 21 E. The small hill, less than 50 feet across and 20 feet high, and surrounded on three sides by swampy ground, is composed of gray, saccharoidal dolomite of semi-biohermal character. A few poorly preserved corals were the only fossils observed. The main part of the hill is covered with large blocks which have separated from the main mass.

The Saukville bioherms appear to be the northernmost ones that are developed in the Racine formation, but they are not, of course, the northernmost of the Niagaran bioherms (Fig. 1).

SUMMARY AND PALEOGEOGRAPHY

In Middle Silurian time eastern Wisconsin was covered by a shallow sea wherein algae, stromatoporoids, corals, crinoids, brachiopods, mollusks (especially cephalopods), and trilobites flourished, building biohermal mounds above the flat calcareous bottoms.

These bioherms were islands of intensive vital development and competition and also sites of extensive precipitation of calcareous matter. They grew from calcareous material produced by the inhabiting organisms and from the salts precipitated directly from the adjacent sea water through physico-chemical means. Almost immediately after their initiation the bioherms became low mounds which stood a few feet to a few tens of feet above the surrounding bottoms, and as they continued to grow, meanwhile maintaining the relations just stated, they gradually took on the cuboidal, domal, or ridgelike form that they exhibit in present exposures.

The surrounding bottoms, which seem never to have been heavily populated, were built up slowly from the fine calcareous materials produced by erosion of the growing bioherms or as precipitates from the overlying water, brought down under favorable physico-chemical conditions in the interbiohermal lagoons.

Today the bioherms appear as irregular bodies of massive and structureless dolomite interrupting the uniform bedding of the regional stratigraphic sequence. They are usually separated from the bedded interbiohermal rocks by a narrow flanking zone across which there is a gradual lithological change from biohermal rock to typical interbiohermal deposits and a sharp structural change from massive biohermal core to horizontally bedded deposits typical of the region. In the latter relationship the horizontal beds upon approaching a bioherm gently rise toward it with increasing dips, and often with increasing thicknesses, until they lose their identity through transition into the core rock. This relation suggests that the growing bioherms were surrounded by talus deposits of clastic calcareous sands, in which the stratification developed quaquaversally from the biohermal mound.

Hills (klintar) owing their preservation largely to the superior resistance of the biohermal core rock were favorite quarry sites in the early days, and amateur collectors soon discovered that the rock in them was highly fossiliferous. As a consequence most of the large collections of Silurian fossils from eastern Wisconsin, with the exception of those made at Racine, are from bioherms, and most of the species listed from the regions where bioherms are known to occur are very likely from core or flank rock. So far as known there has been no recent study of the interbiohermal faunas.

WORKS TO WHICH REFERENCE IS MADE

Alden, W. C. (1906) *Description of the Milwaukee quadrangle, Wisconsin,* U. S. Geol. Survey, Folio 140.

——— (1918) *The Quaternary geology of southeastern Wisconsin with a chapter on the older rock formations,* U. S. Geol. Survey, Prof. Paper 106, 356 pages.

Chamberlin, T. C. (1877) *Geology of eastern Wisconsin* in *Geology of Wisconsin,* vol. 2, Wis. Geol. Survey, p. 91-405.

——— (1883) *General geology:* I. *Chemical geology;* II. *Lithological geology;* III. *Historical geology* in *Geology of Wisconsin,* vol. 1, Wis. Geol. Survey, p. 1-300.

———, and Salisbury, R. D. (1906) *Geology,* vol. 2, *Earth History,* Henry Holt and Company, 692 pages.

Cumings, E. R. (1930) *Silurian reefs near Tiffin, Carey, and Marseilles, Ohio,* Ind. Acad. Sci., Pr., vol. 39, p. 199-204.

——— (1932) *Reefs or bioherms?* Geol Soc. Am., Bull., vol. 43, p. 331-352.

———, and Shrock, R. R. (1927) *The Silurian coral reefs of northern Indiana and their associated strata,* Ind. Acad. Sci., Pr., vol. 36, p. 71-85.

——— — ——— (1928a) *Niagaran coral reefs of Indiana and adjacent States and their stratigraphic relations,* Geol. Soc. Am., Bull., vol. 39, p. 579-620.

——— — ——— (1928b) *The geology of the Silurian rocks of Northern Indiana,* Dept. Conservation, Indiana, Div. Geol., Publ. 75, 226 pages.

Ehlers, G. M. (1921) *Niagaran rocks of the northern peninsula of Michigan* [Abstract], Geol. Soc. Am., Bull., vol. 32, p. 129-130.

Fenton, C. L. (1931) *Niagaran stromatoporoid reefs of the Chicago region,* Am. Midland Nat., vol. 12, p. 203-212.

Grabau, A. W. (1903) *Paleozoic coral reefs,* Geol. Soc. Am., Bull., vol. 14, p. 337-352.

——— (1921) *A Textbook of geology, Pt. II. Historical geology,* D. C. Heath and Company, 976 pages.

——— (1924) *Principles of stratigraphy,* 2d ed., A. G. Seiler, 1185 pages.

Hall, James (1862) *Physical geography and general geology* in *Rept. Geol. Survey, State of Wisconsin,* by Hall, James, and Whitney, J. D., p. 1-72.

Schuchert, C., and Dunbar, C. O. (1933) *A textbook of geology, Pt. II. Historical,* 3d ed., John Wiley and Sons, 551 pages.

——— — ——— (1937) *Outlines of historical geology,* 3d ed., John Wiley and Sons, 268 pages.

Shimer, H. W. (1933) *An introduction to the study of fossils,* Rev. ed., The Macmillan Company, 496 pages.

Shrock, R. R. (1929) *The klintar of the Upper Wabash valley in northern Indiana,* Jour. Geol., vol. 37, p. 17-29.

——— (1935) *Silurian geology of Wisconsin* [Abstract], Geol. Soc. Am., Pr., p. 107-108.

Thwaites, F. T. (1914) *Recent discoveries of "Clinton" iron ore in eastern Wisconsin,* U. S. Geol. Survey, Bull. 540, p. 338-342.

Twenhofel, W. H. (1932) *Treatise on sedimentation,* 2d ed., Williams and Wilkins, 926 pages.

———, and Shrock, R. R. (1935) *Invertebrate paleontology,* McGraw-Hill Book Co., Inc., 511 pages.

Whitfield, R. P. (1882) *Paleontology* in *Geology of Wisconsin,* vol. 4, Wis. Geol. Survey, p. 161-363.

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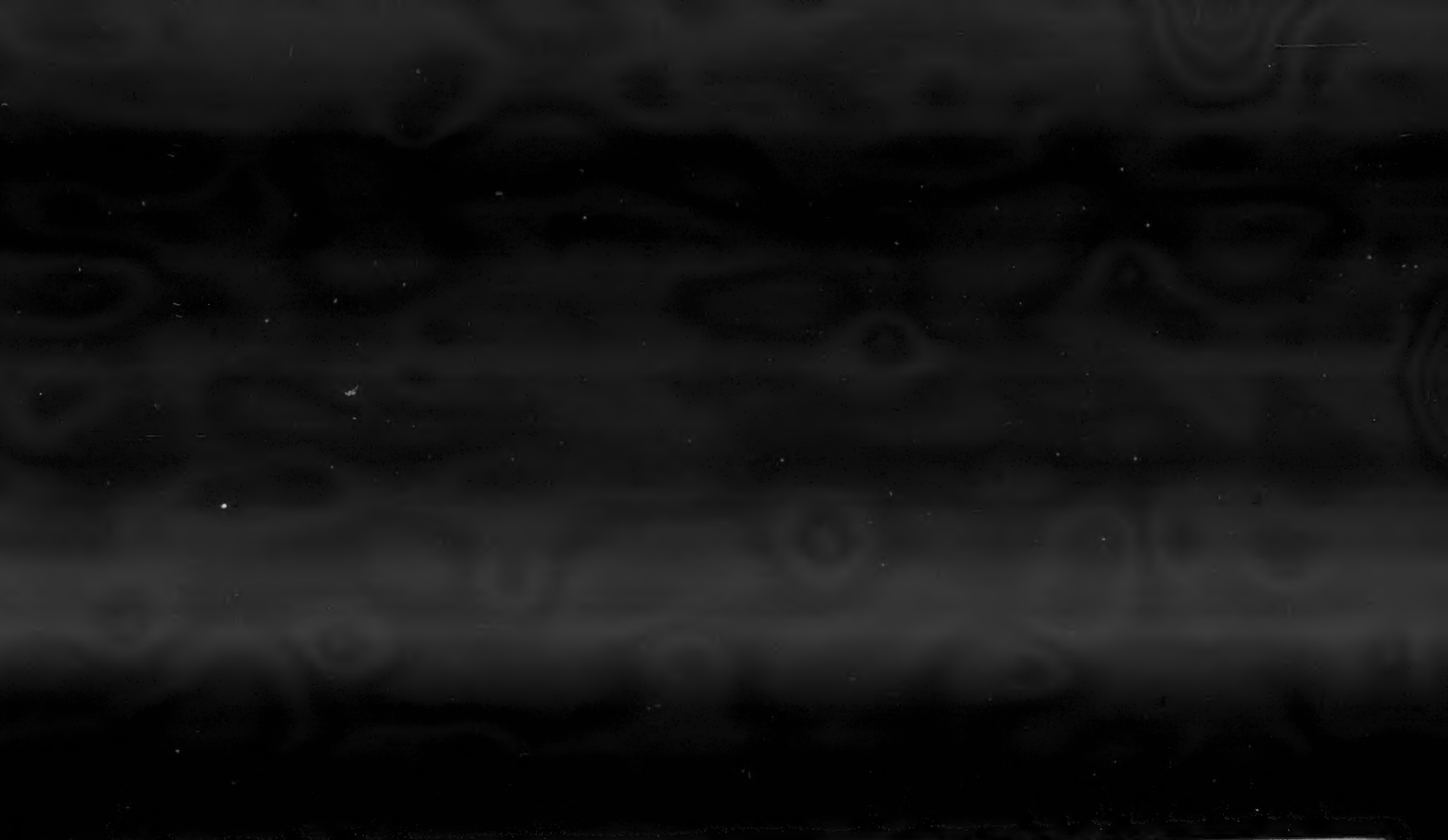
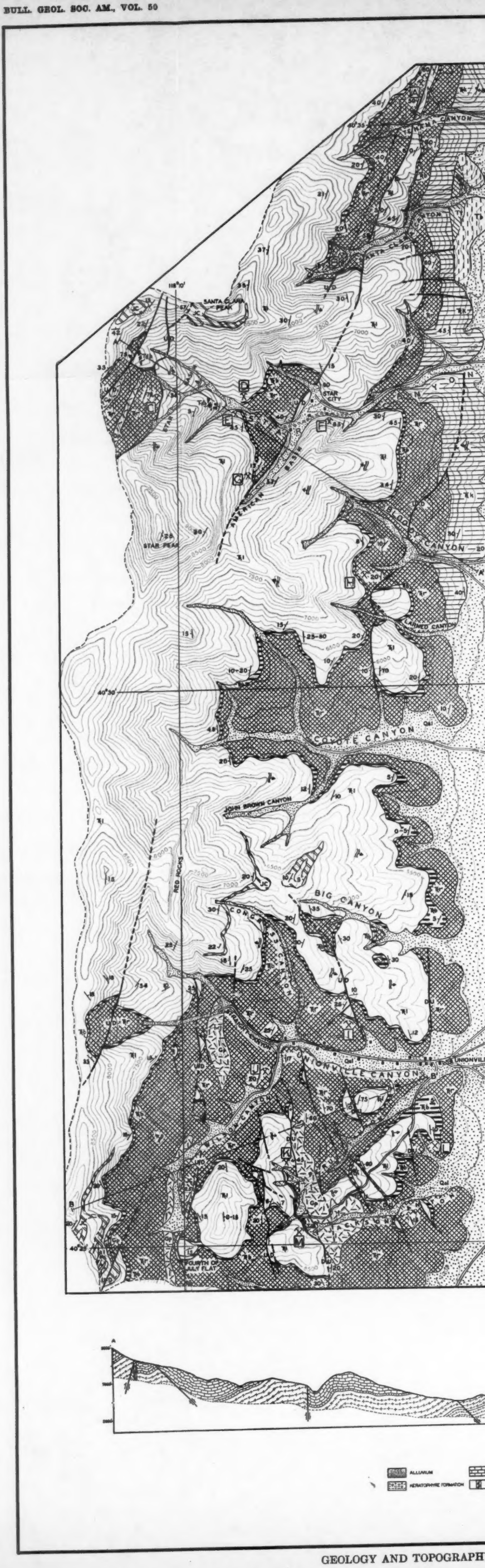



SANTA CLARA PEAK
STAR CITY
STAR PEAK
AMERICAN
BLACK CANYON
LARNED CANYON
COYOTE CANYON
JOHN BROWN CANYON
RED ROCKS
BIG CANYON
UNIONVILLE CANYON
ALLUVIUM

SEDIMENTARY ROCKS

MIDDLE AND UPPER TRIASSIC

- ALLUVIUM — Qal
- STAR PEAK FORMATION
  - UPPER MEMBER — Trl
  - BASAL CLASTICS — Trb

IGNEOUS ROCKS

PLIOCENE (?)

- BASALT — Tb
- DIABASE — Tdb
- TRACHYTE PORPHYRY — Tt

LATE PALEOZOIC, JURA-CRETACEOUS(?) OR TRIASSIC

- GRANITE PORPHYRY — Gp
- METADIABASE META-GABBRO METASYENITE — JC
- RHYOLITE FORMATION — Trp
- KERATOPHYRE FORMATION — Trk

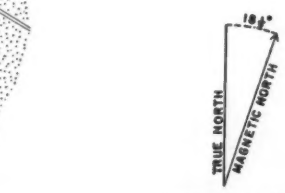
- STRIKE AND DIP OF BEDS — 20
- GENERAL DIP OF CONTORTED BEDS — 4
- FAULTS — U/D, 70

TEHAMA MINE	A
SANTA CLARA MINE	B
CHAMPION MINE	C
SHEBA MINE	D
DE SOTO MINE	E
PFLUM MINE	F
SO. DE SOTO WORKINGS	G
BLOODY CANYON MINE	H
LEAD MINE	I
MARIGOLD MINE	J
ARIZONA MINE	K
BLACK WARRIOR MINE	L
WHEELER MINE	M

NORTHEASTERN HUMBOLDT RANGE

RECONNAISSANCE MAP

E. N. CAMERON



APPROXIMATE MEAN DECLINATION - 1935

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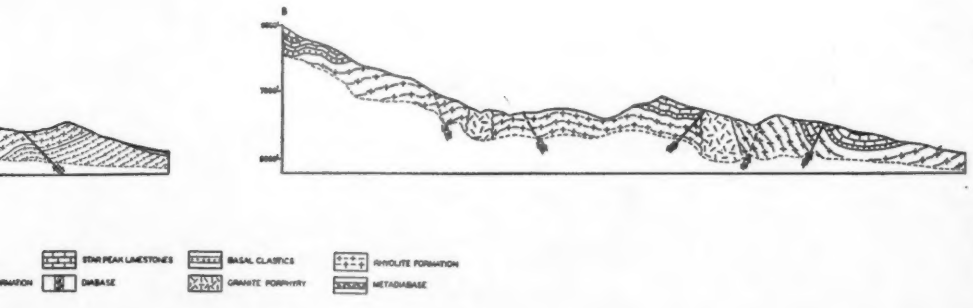

GRAPHY OF THE NORTHEASTERN HUMBOLDT RANGE

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GEOLOGY AND MINERALIZATION OF THE NORTHEASTERN HUMBOLDT RANGE, NEVADA

BY EUGENE N. CAMERON

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ABSTRACT

This paper presents the results of an investigation of the geology and mineralization of the northeastern portion of the Humboldt Range, Nevada. The principal rock units of this area are keratophyric and rhyolitic volcanics overlain with apparent conformity by the Middle and Upper Triassic Star Peak limestones and associated clastic strata. These strata have been folded, faulted, and intruded by Late Jurassic or Early Cretaceous diabase, gabbro, syenite, and granite porphyry. Extensive mineralization and hydrothermal alteration have also taken place. Later deformation and igneous activity, which probably accompanied uplift of the present Humboldt Range by Pliocene "Basin Range" block faulting, are shown by olivine diabase dikes, by post-mineral faulting (in part), and by basalt, trachyte porphyry, and trachytic ash.

The hypogene ore deposits of silver, antimony, and gold are sulphide-bearing quartz fissure veins, bed veins, and stockworks in rhyolite, in granite porphyry, and in the lower Star Peak strata. Hypogene fissure filling and replacement have been accompanied by extensive wall-rock alteration. Supergene deposits comprise portions of the hypogene silver deposits enriched by supergene argentite and native silver. The ore deposits are grouped along major fault zones which trend roughly parallel to Jura-Cretaceous (?) folds and apparently were initiated during Jura-Cretaceous (?) orogeny. The forms of the hypogene deposits and the distribution of the hypogene vein minerals within them show a close control by structural features produced by deformation prior to and during mineralization.

The hypogene ores probably were formed during a single epoch of mineralization and under essentially similar moderate temperature-pressure conditions. Differences among hypogene deposits may be due largely to deformation during mineralization. The Jura-Cretaceous (?) magma was apparently the source of the mineralizing solutions. Supergene enrichment appears to have been essential to formation of workable ores in most of the silver deposits.

INTRODUCTION

GENERAL CONSIDERATIONS

Before the construction of the first transcontinental railroad in 1868 silver mines were in active operation in the northeastern Humboldt Range. The rich ores more than justified transportation 400 miles overland by ox team to San Francisco and shipment thence around Cape Horn to smelters in England and Wales. Considerable excitement attended the discovery and exploration of these silver deposits. The early bonanzas were short lived, however, and general interest in the region quickly subsided. Nevertheless, the mines continued in operation until 1880 and have been worked intermittently since that time. Minor discoveries of antimony and gold have also been made.

The most productive period in the operation of the silver mines came before the era of detailed geological studies in the western United States. By the time such studies became general, the mines had ceased to be important, and the attention of geologists was drawn to more active mining districts. Thus this district, even to the present day, has lacked the geological study which it merits. The present investigation has been undertaken to provide such a study. The purpose has been to map the geology of the area, to visit such parts of the mines as might be accessible, to determine the nature of the mineralization, and to study the relationship of ore deposition to the general geology.

Mining in the northeastern Humboldt Range began with the discovery of silver deposits in Star Canyon in 1861. The area produced three noteworthy silver mines: the Sheba and De Soto mines in Star Canyon (Pl. 1), and the Arizona mine south of Unionville. The Sheba mine was worked for several years, but by 1865 the bonanzas near the surface were exhausted, and this mine and the neighboring De Soto mine were abandoned for a few years. Subsequently, however, intermittent mining was resumed and continued until about 1880. The Arizona mine, discovered in 1862, was worked with few interruptions from about 1865 to 1880. During this period other smaller deposits of antimony, silver, and gold in the area were explored.

The year 1880 seems to have marked the termination of the early period of mining activity in the region, during which the bulk of the production was made. Decline in the value of the ores obtained, together with the

fall in the price of silver attending demonetization of the metal in the 1870's, more than offset decreases in mining costs due to improvements in methods and to cheaper transportation after opening of the Central Pacific Railroad in 1868. Since 1880 various deposits have been worked intermittently, but the early productivity has never been equalled. In August 1937, except for exploratory work in progress at the Marigold mine, there was little activity in the area.

From the fragmentary records available, the total production of the area may be estimated at about $5,000,000 in silver, between $75,000 and $100,000 in antimony, and perhaps $50,000 in gold.

METHODS OF INVESTIGATION

Four months during the summers of 1934 and 1935 were devoted to field study of the district. Since existing maps of the northeastern Humboldt Range were on scales too small for satisfactory presentation of the geologic features, a reconnaissance topographic map on a scale of 5000 feet to the inch, with a contour interval of 100 feet, was prepared by plane-table triangulation. The geology was then mapped on this base. Subsequently the individual ore deposits and adjacent terrains were studied in considerable detail. Field work has been supplemented by microscopical and X-ray study of suites of rocks and ores collected in the field.

ACKNOWLEDGMENTS

The investigation has been made possible largely by the award to the writer of the Samuel Franklin Emmons Memorial Fellowship in Economic Geology for 1935–1936. Through the kindness of Doctor W. F. Foshag and Doctor E. P. Henderson, specimens of ores from the United States National Museum, collected by earlier investigators, have been made available for study.

The writer takes particular pleasure in expressing his sincere appreciation to Doctor Paul F. Kerr, of Columbia University, at whose suggestion the investigation was undertaken, for supervision in the field and for helpful discussion, criticism, and advice during preparation of this report. Doctor Philip Krieger, of Columbia University, has contributed valuable advice and criticism. To Doctor S. J. Shand, of Columbia University, the writer is indebted for encouragement and advice during preparation of the report. The writer also wishes to thank V. H. Booth, A. F. Hagner, and R. J. Holmes for assistance in the field. The many kindnesses extended to the writer during his stay in Unionville by Mr. and Mrs. C. A. Ernst have greatly facilitated the field work. The writer is also indebted to Mr. Ott F. Heizer, of Lovelock, for many courtesies.

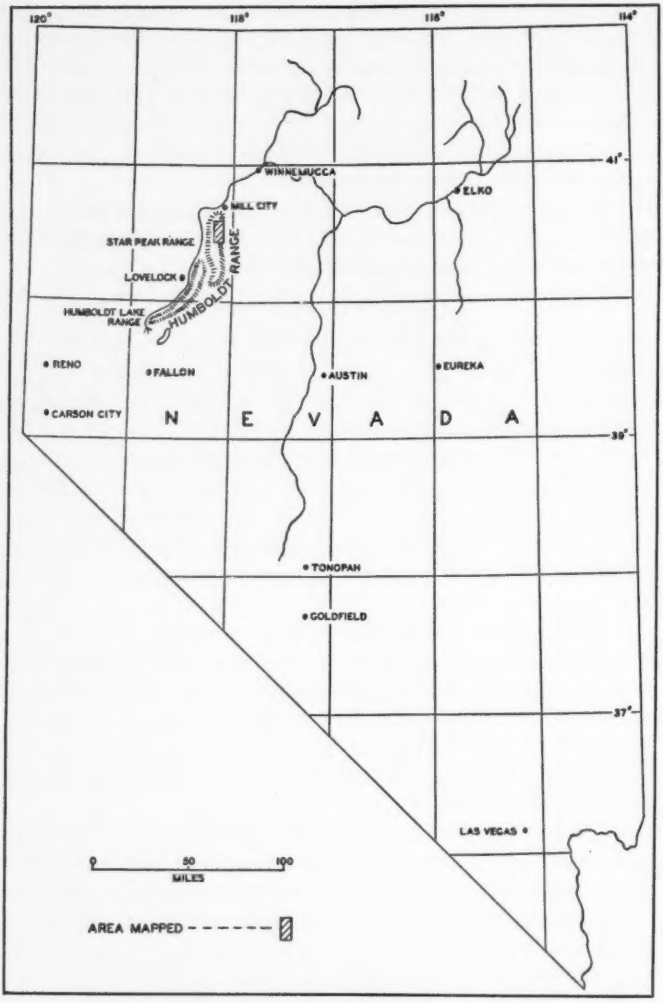


FIGURE 1.—*Index map of Nevada*

LOCATION AND PHYSICAL FEATURES OF AREA

The area studied comprises about 85 square miles on the northeastern slope of the northern division of the Humboldt Range, Pershing County, Nevada, known as the Star Peak Range (Fig. 1). The range has been

studied from Tehama Canyon southward to Cottonwood Canyon, and from the crest to the main road in the Buena Vista Valley. The nearest railroad point is Mill City, about 4 miles north, on the Southern Pacific Railroad. The only settlement is the largely abandoned old mining camp of Unionville, in lower Unionville Canyon at the south end of the area.

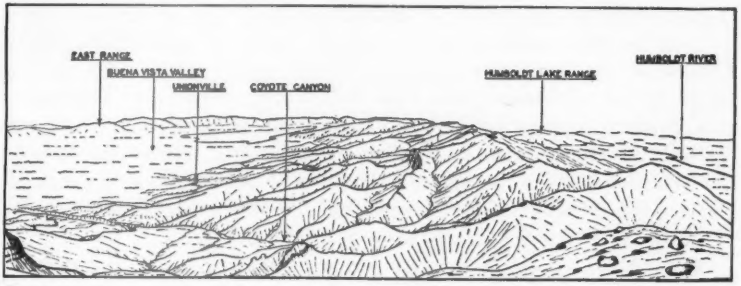


FIGURE 2.—*View southward from Star Peak summit*
Line drawing from photographs.

Where studied, the Star Peak Range has an average width of about 8 miles. Summit elevations along the crest vary from 9835 feet at Star Peak—the highest point—to 7000 feet at the head of Tehama Canyon. Nearly parallel ridges and intervening canyons trend eastward roughly at right angles to the axis of the range (Fig. 2). The ridges slope from the crest to the floor of the Buena Vista Valley, at an elevation of about 4400 feet.

PREVIOUS GEOLOGIC INVESTIGATIONS

Gabb (1864) was probably the first geologist to mention the northeastern Humboldt Range, when he described cephalopods collected in Cottonwood and Unionville canyons. Burton (1868) published analyses of silver-jamesonite from the Sheba mine and freibergite from the De Soto mine. Systematic investigation of the Humboldt Range was begun with the work of the Fortieth Parallel Survey. A reconnaissance of the Star Peak Range was made by Arnold Hague (1877), and the areal geology was mapped on a scale of 4 miles to the inch. The range was described by Hague and King (1878) as consisting chiefly of Triassic and Jurassic rocks folded into a broad anticline trending N. 30° E., oblique to the axis of the range. The Triassic strata were described as 6000 to 7000 feet of metamorphosed sediments, designated the Koipato formation, conformably overlain by an alternation of three great limestone zones and three interposed quartzite zones aggregating nearly 10,000 feet thick, designated the Star Peak formation. Jurassic lime-

stones and slates about 2600 feet thick were stated to overlie the Star Peak formation conformably. J. D. Hague (1870) gave a description of mining activity in the range. A considerable number of Mollusca, chiefly Ammonoidea from the Star Peak formation, were described by Meek (1877). Zirkel (1876) contributed petrographic descriptions of specimens collected from the range. The relation of Pleistocene Lake Lahontan to the Humboldt Range has been described by Russell (1885).

Detailed physiographic and structural studies by Louderback (1904) established the fault-block origin of the Humboldt Range. The granite mass on the west side of the Star Peak Range, considered Archean by the Fortieth Parallel Survey, was recognized as an intrusive and tentatively correlated with late Jurassic intrusives of the Sierra Nevada. The contact between the Star Peak formation and the overlying Jurassic beds in the Humboldt Lake Range (Fig. 1) was reported to be an unconformity[1], although apparently of no great magnitude.

Present knowledge of the invertebrate paleontology of the range is largely due to the important investigations begun by Smith (1904; 1905; 1914; 1927) in 1902. Smith's correlations of the Star Peak formation with other Triassic formations of North America are discussed in a subsequent section. Ichthyosauria from the Triassic limestones have been described by Merriam (1902; 1905; 1908; 1910).

Ransome (1909) made a brief reconnaissance of the region and examined some of the ore deposits of the present area. He recognized the unusual bed veins of the Arizona and neighboring mines and the importance of structural control of ore deposition. He also recognized that a considerable portion of the Koipato formation consists of igneous rocks.

Studies of the Rochester mining district, about 10 miles south of the area, were made by Jones (1913) and Schrader (1914). These were followed by Knopf's (1924) detailed report, which includes a geologic map on a scale of 1:24,000, a classification of the Koipato volcanics, and a detailed discussion of ore deposition. The primary deposits were found to be related to post-Jurassic aplite intrusives. The value of the workable silver ores was found to be attributable largely to supergene enrichment of hypogene protores genetically related to post-Jurassic intrusions of aplite.

The dumortierite deposit in Humboldt Queen Canyon, a short distance north of the Rochester district, was described by Kerr and Jenney (1935). Jenney (1935) also described the geology of the western slope of the Star Peak Range from the Rochester district to a line approximately opposite Jackson Canyon (Pl. 1). The topography and geology

[1] Later confirmed by Smith and Hyatt (Smith, 1905, p. 26).

were mapped on a scale of 2000 feet to the inch. The intrusives of this area and the structural and physiographic history of the Humboldt Range were discussed in detail. Mineralization of the range was ascribed largely to the intrusive granite mass earlier described by Louderback.

Recently Kerr (1938) has described the unusual pegmatitic tungsten mineralization at Oreana, on the west slope of the range. The mineralization has been found to be closely associated with late aplite dikes and masses.

GENERAL GEOLOGIC FEATURES

PRELIMINARY STATEMENT

The oldest rocks in the area are interbedded keratophyric tuffs and flows at least 1500 feet thick, overlain with apparent conformity by interbedded rhyolite tuffs and flows about 900 to 1800 feet thick. The age of the volcanics is uncertain, but according to Wheeler most of them may be late Paleozoic. The entire series is overlain, generally with apparent conformity, by Middle and Upper Triassic limestones and associated clastic strata.

These rocks have been folded and faulted along axes trending north-northeast, slightly oblique to the trend of the present range, and have been invaded by intrusives ranging from olivine diabase to granite porphyry. Deposition of ores of silver, gold, and antimony has been effected by solutions genetically related to the intrusives. This orogeny and its attendant intrusion and mineralization in the Star Peak Range are generally referred to late Jurassic or early Cretaceous time.

Peneplanation followed, but this erosional cycle was interrupted, probably in Pliocene time, by uplift of the present Star Peak Range as a long, narrow, eastward-tilted block, chiefly along normal faults marking the western margin of the range. Uplift took place in two principal stages, separated by an interval during which the range was maturely dissected. Olivine basalt flows now found with trachyte porphyry and trachytic ash along the base of the range in the present area were extruded toward the close of the interval of dissection. Scattered unaltered olivine diabase dikes probably represent fillings of fissures from which olivine basalt flows once issued.

The stratigraphic relationships are summarized in Table 1. The section of the Star Peak formation is that of Smith (1914, p. 8) interpreted on the basis of stratigraphic data disclosed by the present investigation. The relation of the Star Peak formation to the Jurassic, as given by Smith, is included. Jurassic strata, however, have not been found in the area mapped. Thicknesses stated in Table 1 are approximate.

TABLE 1.—*Columnar section—Northeastern Humboldt Range*

Age	Mode of occurrence–stratigraphic divisions		Description
Quaternary	Alluvium and bench gravels.		Gravels, conglomerates, fanglomerates.
Pliocene (?)	Chiefly flows and intrusives.		Olivine basalt, trachyte porphyry, trachytic ash, olivine diabase.
Jura-Cretaceous (?)	Intrusives		Metagabbro porphyry, meta-olivine diabase, metasyenite, and granite porphyry.
Jurassic			Argillaceous slates, carrying *Arietites*.
	Unconformity		
Upper Triassic	Star Peak Formation	*Pseudomonotis* zone (with *P. subcircularis*, *Rhabdoceras*, *Halorites*).	Upper part carbonaceous limestones. Thickness 700 feet, top not exposed. Thin-bedded limestones, with interbedded shales, sandstones, conglomerates. Thickness 500 feet.
Middle Triassic		Coral zone	Massive limestones. Probably corresponds to Hosselkus limestone of California. Thickness 1000 feet.
		Halobia beds	
		Daonella dubia zone.	Thin-bedded limestones, with much sandstone and shale in lower 400 feet. *Ceratites trinodosus* fauna in lower 200 feet. Thickness 1200 feet.
			Basal clastics. Tuffaceous sandstones, shales, and conglomerates. 50 to 200 feet thick.
	Unconformity (?)		
Late Paleozoic or Triassic	Rhyolite formation		Interbedded rhyolitic flows and tuffs. Thickness 900–1800 feet.
	Keratophyre formation.		Interbedded keratophyric flows and tuffs. Thickness at least 1500 feet.

KERATOPHYRE FORMATION

General statement.—The oldest rocks encountered in the area are dark keratophyres and related volcanics which form lines of craggy outcrops in the northern foothills. The true thickness of these rocks is unknown, for, although the top of the formation can be rather closely placed, the

base is not exposed. On the second spur south of Santa Clara Canyon, as indicated on Plate 1, at least 1500 feet of keratophyric rocks can be observed. The belt underlain by the formation along Star Canyon has greater width but cannot be relied upon for stratigraphic thickness, because of possible repetition of strata by strike faulting.

Interest in the problem of spilitic and keratophyric rocks, attested by the voluminous literature which has accumulated since the classic papers of Bailey and Grabham (1909) and Dewey and Flett (1911), warrants a somewhat detailed discussion of the keratophyre formation. Although keratophyres have previously been described from the southern part of the Star Peak Range, in the northeastern part the keratophyric rocks, though invariably altered, are less drastically modified, and the characters of the original rocks can be more accurately inferred.

Description.—Most of the formation consists of dark, aphanitic, massive or flow-banded lavas, in some of which feldspar phenocrysts, dark chloritic spots, or rows of white calcite or quartz amygdules are visible. Some individual flow members are several hundred feet thick, but these probably consist of more than one flow. Intercalated with the flows are green, gray, and dull purple, fine- to coarse-grained tuffs. Bedding, well developed in certain fine-textured members, is lacking in most coarser members.

The lavas include a variety of petrographic types having in common: (1) Meagerly to moderately porphyritic felsitic texture; (2) albite as the prevailing feldspar of both phenocrysts and groundmass; (3) an invariably altered, locally schistose condition. In connection with the problem of the origin of these rocks it is significant that the prevailing flow textures are relict trachytic or pilotaxitic textures which suggest that the now-altered rocks represent in large part original trachytic or andesitic lavas. Relict intersertal texture, suggestive of original basalt, is less common. Since the prevailing lavas are altered trachytic, pilotaxitic, or microfelsitic extrusive rocks consisting essentially of albite and characteristic secondary materials such as chlorite, sericite, epidote, and carbonate, the name keratophyre is applied to the formation. However, the formation also includes spilites and quartz keratophyres. Important types are here described.

A considerable portion of the formation consists of green or dark aphanitic flow-banded keratophyres which in thin section (Pl. 5, fig. 1) show phenocrysts of polysynthetically twinned albite in an altered trachytic to pilotaxitic groundmass consisting chiefly of albite. The albite crystals generally show partial replacement by, and are accompanied by, sericite, with minor amounts of carbonate, zoisite (?), epidote, and chlorite. In some specimens the groundmass texture has been obscured by alteration. Pseudomorphs of various combinations of carbonate, chlorite, epidote,

quartz, magnetite, and sericite are present and in some specimens appear to have been derived from an original amphibole or pyroxene. Accessories are magnetite (chiefly "dust"), apatite, and altered rutile. In part the secondary minerals are introduced, composing veinlets and in some sections amygdules. Partial silicification is shown by a few rocks of this type.

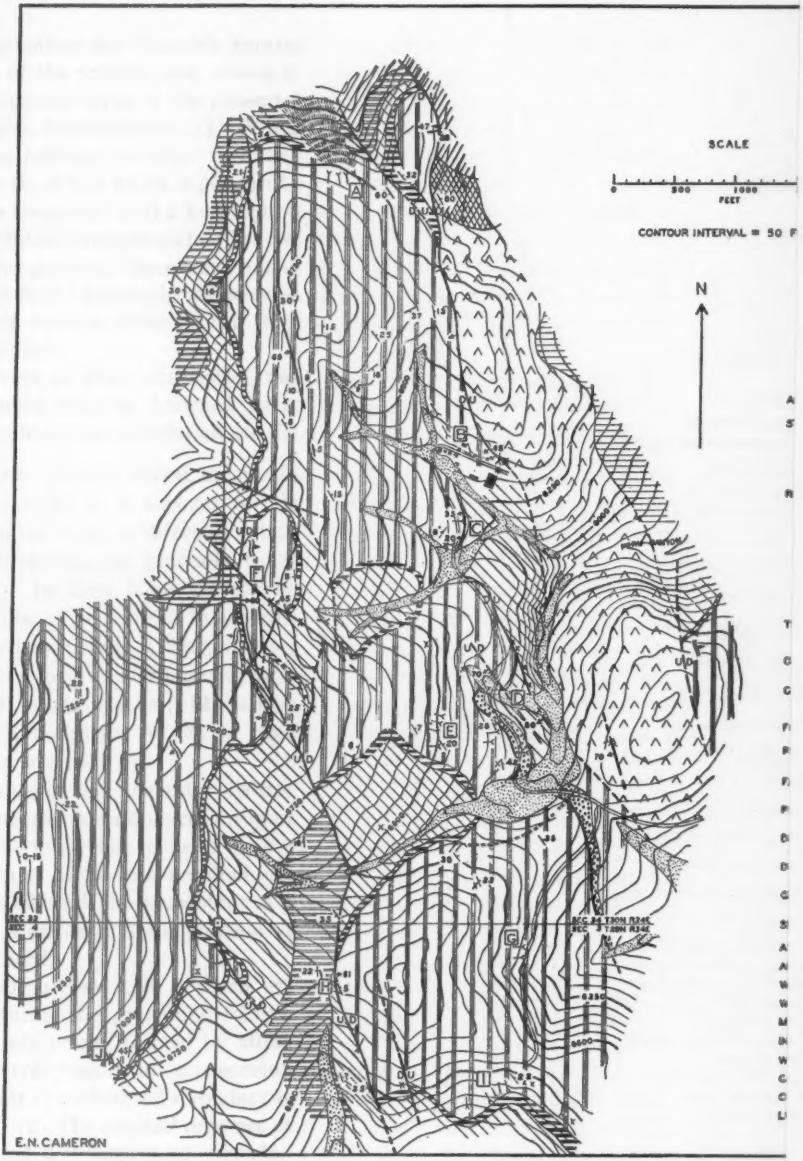
North of Star Canyon considerable areas on the eastern side of the keratophyre outcrop belt are underlain by a green, dense, massive quartz keratophyre with conspicuous dark-green chloritic spots. In thin section the rock shows albite and augite phenocrysts. Many of the albite phenocrysts have been replaced by mixtures of sericite, epidote, chlorite, and carbonate. The augite phenocrysts have been partly replaced by chlorite, epidote, and carbonate. The felsitic groundmass consists of rectangular crystals of alkali feldspar (apparently largely albite) and pyroxene, partly altered in the same manner as the phenocrysts. Interstitial quartz, apparently largely primary, is present. Apatite, altered sphene, zircon, and magnetite are accessories. In part the secondary minerals have been introduced, forming veinlets.

A prominent member of the formation in the lower foothills between Bloody Canyon and Star Canyon is a greenish-gray, aphanitic rock containing phenocrysts of altered feldspar. This rock differs chiefly from the rocks already described in containing unmistakable pseudomorphs after olivine phenocrysts, in a groundmass exhibiting intersertal or a rude fluidal texture. This rock appears definitely to be an altered olivine basalt, which may be classified as a spilite.

The intercalated tuffs examined in thin section consist of abundant angular lithic fragments of types similar to the enclosing flows, embedded in fine-textured matrix material (Pl. 5, fig. 2). In two specimens the matrix material shows relict vitroclastic textures, although the original glass is altered to sericite and fine quartz. In most of the tuffs examined alteration similar to that exhibited by the lavas has obscured primary matrix structures, but their pyroclastic origin is inferred from field observations and from their similarity to the known tuffs.

The keratophyric rocks are sheared and in places schistose. Schistosity is especially well developed in the tuffs, particularly in Santa Clara and Tehama canyons. Locally, however, shearing of lavas has produced rocks which consist of unreduced "augen" of lava in fine schistose sericitic and chloritic material. These strikingly resemble sheared coarse tuffs.

Correlation of keratophyre formation.—A series of highly altered keratophyric rocks in the Rochester district has been described by Knopf (1924, p. 20-23). More recently Jenney (1935, p. 19-23, and map) has named



MAP OF VICINITY OF ARIZONA GROUP OF DEPOS

Pace-compass survey based on triangulation by R. V. Thornl

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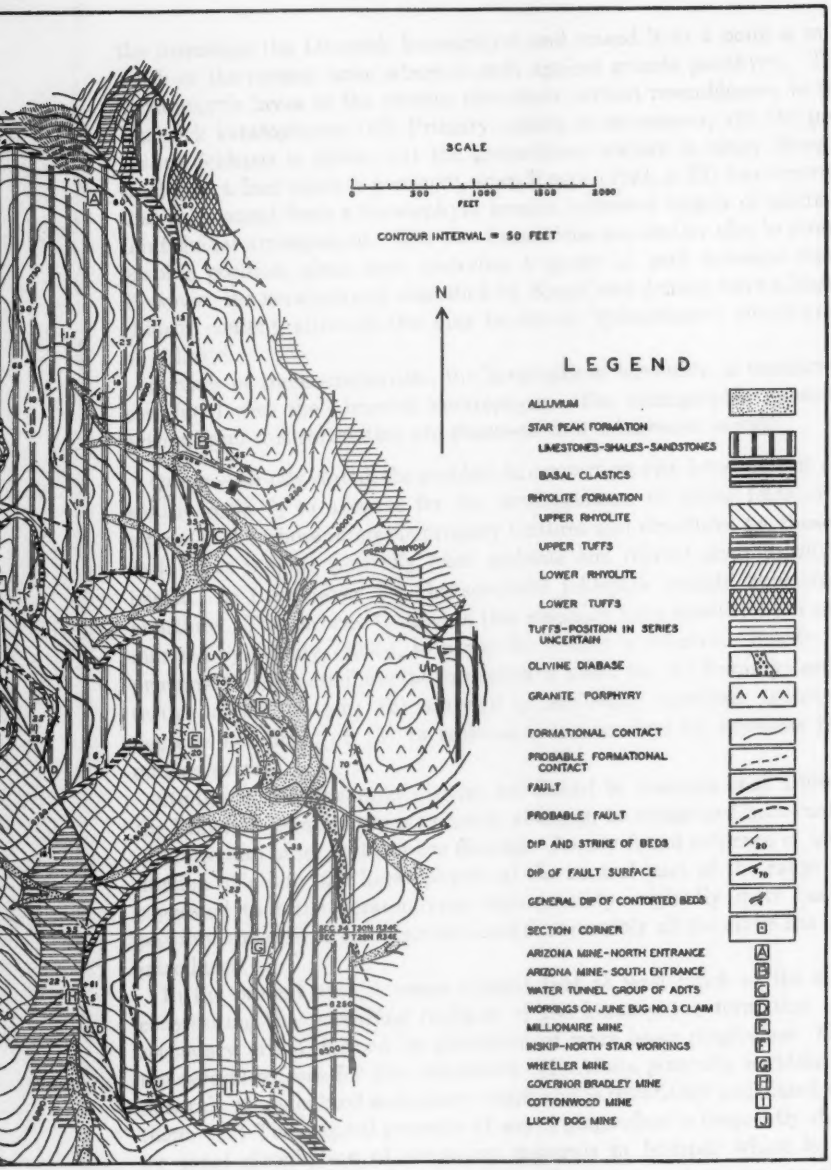

MAP OF VICINITY OF ARIZONA GROUP OF DEPOSITS

Pace-compass survey based on triangulation by R. V. Thornburgh.

the formation the Limerick keratophyre and traced it to a point a mile south of the present area, where it ends against granite porphyry. The keratophyric lavas of the present area show certain resemblances to the Limerick keratophyres: (1) Primary quartz is uncommon; (2) the prevailing feldspar is albite; (3) the groundmass texture in many flows is trachytic, a fact which is pertinent, since Knopf (1924, p. 21) has reported that a fragment from a keratophyre breccia consisted largely of albite in faint fluidal arrangement. The two formations are similar also in stratigraphic position, since each underlies a group of acid volcanic rocks. However, the keratophyres described by Knopf and Jenney have a higher chlorite content, although this may be due to hydrothermal introduction of chlorite.

In view of these similarities, the keratophyre formation is tentatively correlated with the Limerick keratophyre. The stratigraphic questions arising from this correlation are discussed in a subsequent section.

Origin.—Briefly stated, the problem in connection with keratophyric and spilitic rocks is to account for the predominance of albite feldspar in altered rocks whose inferable primary textures and structures are those of rock types ranging generally from andesite and related lavas to olivine basalt. In these lavas the characteristic feldspars contain appreciable anorthite. The theories to explain this anomaly have recently been summarized by Gilluly (1935), to whom the reader is referred. Briefly, the principal theories attribute the formation of albite to: (1) Primary (orthotectic) crystallization; (2) activity of end-stage solutions (autolysis, deuteric activity); or (3) to regional metasomatism by solutions from outside sources.

Studies in the Rochester district led Knopf to conclude that albite in the Limerick keratophyre is primary, although he recognized introduction of albite into other rocks of the district. Jenney found evidence of introduction of albite into keratophyres of the central part of the range and suggested that the keratophyres were possibly originally more basic—possibly andesitic in composition—and that possibly all the albite has been introduced.

In the present area it seems evident that at least much of the albite constituting the prevailing feldspar of the keratophyre formation is of secondary origin, formed by alteration of more basic plagioclase. There are several reasons for this conclusion: (1) Albite generally is riddled by minerals of undoubted secondary origin and is invariably associated with them. (2) The original presence of zoned plagioclase is frequently shown by zonal distribution of secondary minerals in feldspar which is now entirely albite. (3) The prevailing relict primary textures are those

common in andesites and trachytes, in which the usual plagioclase feldspars are anorthite-bearing varieties; also, in a few specimens there are relict textures and minerals of olivine basalt. Thus the nature of the original rocks inferable from textural features is in accord with the conception, suggested by the condition of the albite and its intimate association with secondary minerals, that the albite is an alteration product of more basic plagioclase. (4) The type of albitization in the keratophyres is shown also by plagioclase in Jura-Cretaceous (?) metadiabase and metagabbro, and in the metagabbro all stages in the alteration of andesine-labradorite to albite and associated secondary minerals can be traced.

There is no evidence, either in the intrusives or in the keratophyres, that albitization is due to introduction of albite. In the southern part of the area, where introduction of albite into rhyolite tuffs and granite porphyry is known to have taken place, the process is clearly indicated by veinlets of quartz-albite material. The absence of such criteria in the keratophyric and basic rocks suggests that albitization is the result of some process other than introduction of albite. The lack of these same criteria, together with the widespread nature of this type of alteration and the fact that it has affected rocks so diverse both in age and original composition, also suggests that albitization is not due to any such local process as autolysis (Bailey and Grabham, 1909, p. 252) or deuteric activity (Daly, 1933, p. 421).

The development of the sericite, fine quartz, chlorite, and other secondary minerals invariably associated with albite in the keratophyric rocks is traceable to a combination of shearing and hydrothermal alteration. Albite appears to have formed at the same time and by the same processes as these minerals. Together with albite, these minerals constitute a kind of assemblage which is frequently produced by the type of low-grade metamorphism and alteration which is known to have attended orogeny and mineralization in the area. It is therefore concluded that the "albitization" and accompanying alteration of the keratophyric rocks and the Jura-Cretaceous (?) basic rocks is the result of Jura-Cretaceous (?) shearing and hydrothermal alteration. The change from anorthite-bearing feldspar to albite has generally taken place metasomatically by simultaneous abstraction of calcium from the feldspar molecule and formation of calcite, epidote, sericite, and other secondary minerals.

RHYOLITE FORMATION

General description.—Interbedded rhyolitic flows and pyroclastics underlie large portions of the area. In places exposures form prominent cliffs or craggy outcrops (Pl. 3, fig. 1). Deposits of silver, gold, and antimony are found in the upper part of the formation.

The lavas are gray, light-colored, or purple, porphyritic, blocky felsites; outcrops are stained shades of brown by limonite. Banded or streaky flow structure is characteristic, and some of the flows have a slabby or thinly platy parting (Pl. 3, fig. 2) parallel to flow structure. Phenocrysts of feldspar are usually conspicuous in hand specimens, though generally small and sparsely distributed. Quartz phenocrysts are seldom numerous, but quartz streaks and spots representing vesicle fillings are abundant. Individual flows range from 20 to 75 feet thick. Thick members are probably multiple flows.

The pyroclastic strata range from coarse-grained lithic or crystal tuffs with fragments a foot or more in diameter to extremely fine textured ash beds. The ash beds in places (Pl. 3, fig. 3) show thin, remarkably even bedding, but many coarser tuffs are without apparent bedding.

Thickness.—The rhyolite formation at Bloody Canyon is approximately 1800 feet thick. The thickness decreases northward and is only about 900 feet at Tehama Canyon. In Wilson Canyon about 1500 feet of rhyolitic volcanics is present, but here the base of the formation is not exposed.

Petrography.—Despite extensive alteration a great variety of obsidians, perlites, rhyolitic vitrophyres, and porphyritic rhyolites are still recognizable among the lavas. In porphyritic types albite phenocrysts ($n\alpha = 1.526\text{-}1.528$; $n\gamma = 1.536\text{-}1.538$), in part of the "chessboard" variety, are the most abundant. Orthoclase (soda orthoclase) phenocrysts are less common. Quartz phenocrysts are present in some flows. Some lavas contain scattered pseudomorphs, apparently derived from biotite phenocrysts, consisting of chlorite, secondary magnetite, bleached mica, leucoxene, and sagenite webs. The essential groundmass constituents are orthoclase and quartz, accompanied in some flows by albite. Orthoclase also rims, corrodes, and partly replaces albite phenocrysts in many of the flows. Small grains of apatite, zircon, magnetite, and sphene are minor accessories. In thin sections the prevailing flow textures are seen to be due to alternation of finer with coarser microspherulitic bands (Pl. 5, fig. 3), or with microfelsitic bands; to parallel alignment of phenocrysts; or to elongate cavities filled with quartz or lined with orthoclase or albite crystals. The spherulites consist of orthoclase or of orthoclase and quartz.

The coarser clastic strata examined in thin section consist of abundant crystal fragments, chiefly quartz and alkali feldspars, and angular lithic fragments of acid lavas similar in general to the lava flows. The matrix material in all specimens is highly altered, consisting essentially (Pl. 5, fig. 4) of fine mixtures of sericite, quartz, and feldspar, and, in a few specimens, small amounts of chlorite. The fine-textured members are composed entirely of such mixtures. Quartz-albite mixtures form veinlets in tuffs

near the Arizona mine. In view of the generally altered condition of the pyroclastics, it is significant that clearly defined relict vitroclastic structure was found in two of the specimens examined in thin section—one from the south fork of Bloody Canyon, the other from beds near the Arizona mine. The structure confirms the pyroclastic origin inferred for these beds from their field characters and suggests that the rhyolitic clastics, at least in large part, represent accumulations of pyroclastic material. However, the remarkably even lamination shown by some of the ash beds suggests deposition in water.

Like the keratophyres, the rhyolitic volcanics have been sheared and hydrothermally altered, chiefly by sericitization and silicification; locally flows and tuffs are indistinguishable. Oxidation of disseminated introduced pyrite is the principal cause of the brown colors of outcrops. Particularly near fault zones, the volcanics have been converted to glossy quartz phyllites. Foliation in any given locality is best developed in pyroclastics (Pl. 3, fig. 4). Part of the sericite in the rhyolitic volcanics, however, is due to hydrothermal alteration, for indications of shearing are lacking in certain sericitized specimens.

All specimens examined contain introduced quartz, which has replaced phenocrysts and groundmass minerals (except quartz), filled cavities, and formed veinlets. Parallel to the flow banding in the uppermost flow at Bloody Canyon, and elsewhere, remarkable strings of banded chalcedonic quartz nodules fill primary vesicles. The presence of introduced quartz precludes estimation of primary quartz content in most of the flows. The rocks are classified as rhyolites, however, for quartz phenocrysts are present in some flows, and the primary textures and structures preserved are usually those typical of rhyolites.

STRATIGRAPHIC RELATIONS AND CORRELATION OF VOLCANICS

The volcanics overlying the Limerick keratophyre proper south of the present area were divided by Knopf (1924, p. 14-18, 22-29) into two principal units, the lower designated the Rochester trachyte, the upper the Weaver rhyolite. Jenney (1935, p. 23-29) accepted this division but disagreed with Knopf's classification of the Rochester volcanics; he renamed the formation the Rochester rhyolite. The total thickness of the Weaver and Rochester formations was given by Jenney (1935, p. 18) as 7600 to 8600 feet. This thickness includes 2000 to 3000 feet of interbedded rhyolite and keratophyre underlying the Rochester formation. In the area mapped by Jenney (1935, p. 27), the Weaver and Rochester formations are apparently lithologically similar. The principal reason for separating the two formations was the presence of rhyolite tuffs mapped as the base of the Weaver formation; these tuffs are petrographi-

FIGURE 1. RHYOLITE OUTCROPS
Near Governor Bradley mine.

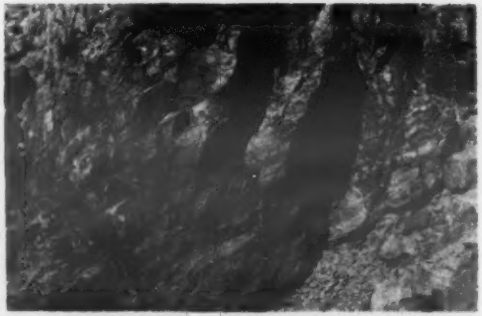

FIGURE 2. NEAR VIEW OF RHYOLITE OF FIGURE 1
Rhyolite shows parting parallel to flow banding, which dips to the right.

FIGURE 3. RHYOLITIC TUFFS
Unionville Canyon.

FIGURE 4. SCHISTOSE RHYOLITIC TUFFS
South side Wilson Canyon.

PHOTOGRAPHS OF RHYOLITIC VOLCANICS

FIGURE 1. TUFFACEOUS BASAL CLASTICS
Near Cottonwood mine. Rock face perpendicular to bedding.

FIGURE 2. TUFFACEOUS BASAL CLASTICS
Joint surface parallel to bedding.

FIGURE 3. THIN-BEDDED STRATA OF LOWEST LIMESTONE MEMBER OF STAR PEAK FORMATION
Hill northeast of Fourth of July Flat.

FIGURE 4. FRACTURE CLEAVAGE
In thin-bedded limestones. Hill northeast of Fourth of July Flat.

PHOTOGRAPHS OF STAR PEAK SEDIMENTS

cally indistinguishable from pyroclastic layers within the rhyolite both above in the Weaver and below in the Rochester (Jenney, 1935, p. 26). That the present writer has not distinguished two groups of acid volcanics in the present area may be due to the similarity described by Jenney. The formational boundary which he mapped ends against granite porphyry several miles south of the present area and consequently cannot be traced into the area. No significant unconformity has been recognized in the volcanics of the present area. Rhyolite and keratophyre formations in the walls of Bloody Canyon and Star Canyon are apparently conformable, although the actual contact is not visible. The zone of interbedded keratophyres and rhyolites described by Jenney and Knopf has not been found.

The keratophyre outcrops in the present area are about 8 miles from the nearest exposures of the Limerick keratophyre mapped by Jenney, no allowance being made for folding and faulting. In this distance, if the tentative correlation made in this report is correct, the apparent thinning of the volcanics overlying the keratophyre formation is about 6000 feet.

The volcanics were assigned to the Triassic in the Fortieth Parallel Survey reports, on the basis of their apparent conformity to the overlying Star Peak formation. Knopf (1924, p. 30-31) described rhyolitic and andesitic volcanics as interbedded with the lower Star Peak strata east of the Rochester district. He concluded that these volcanics represent a continuation of volcanism which earlier in the Middle Triassic produced the Weaver and older volcanics. However, Wheeler (1938) holds that a recently discovered specimen of *Helicoprion* (Wheeler, 1936) requires assignment of the Rochester formation to the Anthracolithic (Uralian or Artinskian) and the Limerick keratophyre to the late Paleozoic.[2] Wheeler further states that the contact of the Weaver and Star Peak formations is a disconformity, that the lower part of the Star Peak formation contains no volcanics, and that consequently the Weaver should likewise tentatively be assigned to the Anthracolithic. In the present area it is possible that a significant disconformity exists at the base of the Star Peak formation, and the thinning of the rhyolites northward into the present area conceivably may be due to pre-Star Peak erosion. In view of the doubt which therefore exists concerning the earlier assignment of the pre-Star Peak volcanics to the Triassic, in the present paper the keratophyre and rhyolite formations are tentatively classified as either Upper Paleozoic or Triassic.

The discovery of an authentic late Paleozoic fossil from the pre-Star Peak volcanics is of unusual interest, since it indicates that Late Paleo-

[2] In this connection it is of interest that Gilluly (1935) has described Permian keratophyres from the Wallowa and Blue Mountains of eastern Oregon.

zoic formations are much more widely distributed in northern Nevada than has previously been recognized.

STAR PEAK FORMATION

General statement.—The most widespread strata mapped belong to the Star Peak formation, largely composed of limestones. Exposures occur chiefly high in the range along the western side of the area, but the lower strata of the formation extend into the foothills. To the southeast, the lower members form a relatively thin veneer over the rhyolites, which are extensively exposed in the deeper canyons. The great cliffs at the head of Congress Canyon and on Star and Santa Clara peaks mark outcrops of thick, massive limestones. The thin-bedded strata generally produce more subdued topographic forms.

The Star Peak formation is economically significant, since it contains the most important mineral deposits of the area.

Subdivisions.—The basal member of the formation comprises readily distinguishable tuffaceous clastics, mapped as a unit. Four lithologic units in the remainder of the formation are here described, but no attempt has been made to map them separately. Determination of structure in much of the formation, owing to its uniformity and to an absence of readily recognizable key beds, was not feasible on the scale of the present study.

Basal clastics.—The basal clastics (Pl. 4, figs. 1 and 2) typically consist of gray, green, brown, or purple tuffaceous sandstones and conglomerates but exhibit a wide range of textural types. Bedding is best developed in the finer strata; in places it is remarkably fine, even, and regular. The thickness of the basal clastics ranges from 50 to 200 feet. This variation is probably due largely to lensing out of beds.

Both in field and microscopic characters, the basal clastics in general resemble the tuffs of the rhyolite formation and may consist in part of pyroclastic material. Shearing and alteration shown are similar. Vitroclastic textures, however, are absent in thin sections.

In Star Canyon and Bloody Canyon the basal clastics are interbedded with thin limestone and shale beds of the Star Peak formation proper. The basal clastics in general appear concordant in attitude with the underlying rhyolite. In Bloody Canyon, however, the strata contain fragments clearly derived from the underlying uppermost rhyolite flow. The magnitude of the disconformity indicated is unknown.

Limestone members.—The great series of calcareous strata constituting the bulk of the Star Peak formation is best exposed on the south side of Star Canyon. This is the type section of the formation (A. Hague, 1877,

p. 723-727; King, 1878, p. 273-277). King gave the thickness of the formation as nearly 10,000 feet, but he overlooked repetition of the beds by strike faulting.

In 1905 Smith and Hyatt (1905, p. 21), referring to King's section, remarked: "No such thickness of limestone is visible in the region, for faulting has caused the same beds to be counted more than once in making the estimate." In 1914, however, Smith (1914, p. 8) published a stratigraphic column for the Star Peak formation which lithologically differs chiefly from that of King in that King's uppermost quartzite is omitted and the presence of volcanics is reported. No reason was given for changing the opinion expressed in 1905.

The present work has shown that strike faulting does exist in the Star Canyon section. King's quartzite No. 1 is the top of the rhyolite formation repeated by the American Basin fault, and the supposedly overlying limestone No. 2 is limestone No. 1 repeated. Similarly, the complex at the head of Star Basin, which consists principally of the top of the rhyolite formation repeated by faulting, seems to have been considered as an "interposed quartzite zone". Although King's type section of the formation, considered in its entirety, therefore does not give an uninterrupted stratigraphic sequence, the portion of his section extending from the top of the rhyolite formation in American Basin to the summit of Star Peak presents an essentially unbroken sequence of the limestones. In this sequence—the most complete afforded by the area—the writer has recognized four lithologic units above the basal clastics.

Limestone No. 1, the lowest member, is about 1200 feet thick. The lower 400 feet, which contains the principal silver deposits of the area, is characterized by gray to blue-black, thin-bedded, finely crystalline, relatively pure limestones (Pl. 4, fig. 3) interbedded with brown, gray, pink, and purplish shaly, silty, and sandy layers, some of which are dolomitic. The principal impurities are chert and detrital quartz, muscovite, and alkali feldspars. Between 100 and 150 feet above the base is a prominent facies, about 100 feet thick, composed of gray, medium- to coarse-grained, dolomitic, limonitic quartzites and quartz sandstones characterized by a rough slabby parting at an angle to the bedding. These strata weather to dark brown slabs which form conspicuous talus slopes at many places in the area.

The upper 800 feet is chiefly gray to blue-black, finely crystalline (dense in hand specimen) limestones, thin-bedded for the most part but containing thick-bedded, massive limestones at a number of horizons. Gash veinlets and irregular patches of coarse white calcite are locally conspicuous.

Overlying limestone No. 1 is 1000 feet of light-gray, finely crystalline,

massively bedded limestone, which forms the great limestone cliffs of the area. The strata are in most places severely fractured, and bedding is indistinct. Ramifying veinlets of coarse white calcite are abundant. At a number of places, as in the "Red Rocks" at the head of Congress Canyon, these limestones contain striking irregular patches colored red or buff by iron oxides derived from disseminated pyrite. Locally caverns have developed in the beds.

The shoulder of Star Peak above the cliffs is composed of the strata of member 3, approximately 500 feet thick. These are gray to bluish-gray, finely crystalline, thin-bedded limestones interbedded with brown and gray shale, sandstones, and quartzite pebble conglomerates, some of which are limonitic.

The topmost member, exposed on Star Peak summit, is composed of dark bluish-gray, carbonaceous, finely crystalline limestone, about 700 feet thick, in beds a few inches to about 2 feet thick. The beds are much fractured and strikingly seamed by veinlets of coarse white calcite. These beds are equivalent to the lower part of King's "Limestone No. 5", described as dipping westward from the summit (King, 1878, p. 277). The actual dip at the summit, however, is to the east.

Shearing and alteration.—Like the underlying volcanics the limestones were sheared during the Jura-Cretaceous (?) orogeny. Near some of the fault zones the silty and argillaceous members have been converted to sericitic calcareous phyllites. In many places fracture cleavage is shown by the thin-bedded limestones (Pl. 4, fig. 4). The more massive limestones have generally yielded chiefly by fracturing, which where severe has destroyed bedding. In thin sections of the limestones deformation is seen to have caused recrystallization of calcite and, in the impure varieties, recrystallization of quartz and development of sericite.

Except near the ore deposits hydrothermal alteration is rarely conspicuous. Quartz veins along fractures in limestone in part may be of hydrothermal origin. Locally the limestones contain conspicuous gash veinlets and spots of coarse white calcite. Chert bands and nodules are frequently prominent, and thin sections of several limestone specimens show partial replacement by quartz of the "jasperoid" variety. Some of the limestones contain disseminated pyrite, but in general no relation between these effects and hydrothermal activity in the range has been established.

Age and correlation.— Certain beds in the lower 200 feet of the limestones are richly fossiliferous. Near the Lucky Dog mine and on the slopes east of the Inskip mine (Pl. 2) excellent collections were obtained from these beds by Smith (1905, p. 22). From the latter locality and

from the spur immediately southwest of the Arizona mill, the writer collected fossils which have been studied by Sperb (1937, p. 8-10), who reports the following species:

Daonella dubia (Gabb)
D. moussoni plana (n. var.)
D. americana mollis (n. var.)
D. rugosa (n. sp.)
Postaumnia auris (n. gen. and n. sp.)
Discanites perplanus (n. gen.)
Lecanites fragilis (n. sp.)
L. modus (n. sp.)
Trachyceras (*Protrachyceras*) *meeki* Mojsisovics
T. (*Protrachyceras*) *senicircus* (n. sp.)
Ceratites magnus-spinus (n. sp.)
Orthoceras campanile Mojsisovics
Beyrichtites rotelliforme (Meek)
Celtites rotundus (n. sp.)
Lecanites parvus (Smith)
Anatropites pilgrimii Diener
Ceratites humboldtensis Hyatt and Smith
Perumtites bicanalis (n. gen. and n. sp.)
Ceratites humboldtensis trini (n. var.)
C. meeki Mojsisovics
C. altilis Smith
C. tenuspiralis Smith
C. russelli Smith
C. trojanus Smith
C. karpinskyii Smith
C. occidentales Smith
Balatonites undatus (n. sp.)
Nevadites sp.

The *Daonella dubia* zone is thus shown to be included in the beds. (*See* Table 1.) Although fossils have been observed at several higher horizons, identifiable specimens have not been obtained, owing to the fractured condition of the strata.

The age of the Star Peak limestones is given by Smith as Middle and Upper Triassic (Table 1). In the classic paleontological studies of Smith, however, fossil localities were not described in sufficient detail to permit exact correlation of his zones with the lithologic units described here, and there are discrepancies between the sections given by Smith (1914, p. 8; 1927, p. 2) in 1914 and 1927. In addition, detailed mapping has disclosed structural features not appreciated in the early work. However, limestone No. 2 of the present section is probably the massive limestone correlated with the Hosselkus limestone of California, and limestones Nos. 3 and 4 probably correspond roughly to Smith's *Pseudomonotis* zone. Although the top of the Star Peak formation is not exposed in the mapped area, it is believed that the formation can be satisfactorily redefined as comprising the Middle and Upper Triassic strata overlying the rhyolite formation.

Thickness.—According to King, west of Star Peak approximately 3000 feet of beds overlies the limestones of member No. 4 as here designated. However, the uppermost member of King's section—2000 feet of quartzites—is omitted from the section given by Smith (1927, p. 2). Smith thus requires a reduction of King's estimate to approximately 1000 feet. Although further study of this uppermost portion of the formation would thus seem to be required, the total thickness of the Star Peak formation may for the present be estimated provisionally to be between 4500 feet and 6500 feet, with the lower figure the more probable.

JURA-CRETACEOUS (?) INTRUSIVES

General considerations.—The Star Peak formation and the older volcanics of the Star Peak Range have been invaded by various intrusives. Previous investigators have recognized two groups: (1) An older group consisting, in order of intrusion, of metadiorite, early aplite, granite porphyry, granite, late aplite, and pegmatite; (2) olivine diabase dikes of probable Pliocene age.

Of the older group, all types except late aplite and pegmatite have been found by Knopf (1924, p. 31-35, and Pl. 1) and Jenney (1935, p. 30-47, and map) to compose intrusions of considerable size, which in the aggregate underlie a major portion of the west slope of the range south of the present area. The formation of hypogene mineral deposits in the range has been attributed to emanations genetically related to the granitic members of the group. It is generally accepted that emplacement of the bulk of the older intrusives attended folding and faulting of the older rocks, probably in late Jurassic or early Cretaceous time, and preceded mineralization. The metadiorite, however, owing to its metamorphosed condition, was tentatively considered by Knopf (1924, p. 31) to "have been intruded before the post-Triassic deformation and therefore during the Triassic igneous activity." Jenney (1935, p. 30-34, 41) described it as the oldest member of the Jura-Cretaceous (?) sequence.

The present area shows a marked contrast to the southern part of the range, for the group of older intrusives is represented here only by granite porphyry. However, the area exhibits intrusive masses of gabbro porphyry, olivine diabase, and syenite which have been affected by the Jura-Cretaceous (?) deformation and attendant hydrothermal alteration and which evidently belong, therefore, to the older group. The relations of these intrusives to one another are not shown, but these rocks are probably genetically closely related to the metadiorite. Since these intrusives are everywhere altered, they are here designated as metagabbro porphyry, meta-olivine diabase, and metasyenite.

Meta-olivine diabase.—Meta-olivine diabase is found at two places in the southwestern part of the area. At Fourth of July Flat an aphanitic green rock forms a dike about 75 feet wide along a fault zone. The rock is an olivine basalt in which olivine phenocrysts have been altered to chlorite, epidote, and carbonate, and plagioclase phenocrysts have been altered to albite and carbonate. The groundmass consists largely of actinolite, presumably derived from pyroxene, and albite. Epidote and carbonate compose veinlets.

Altered greenish-gray olivine diabase (and olivine basalt) underlying a small area at the range crest at the head of Wilson Canyon has been

described by Jenney (1935, p. 47-48). Jenney tentatively correlated these rocks with late Pliocene (?) basalts. However, the extensions of the mass into the present area clearly form a sheet roughly concordant with the enclosing limestone. The mass is therefore either a sill or an interbedded lava. Marginal relations are not exposed, but at one point a 9-inch layer of limestone enclosed in the sheet has been recrystallized to fine-grained marble. This feature, together with the even and parallel contacts of the enclosure, suggests a xenolith rather than a bed deposited between successive eruptions. The alternation of sedimentary strata with igneous bands along the summit is interpreted as branching of the sheet at its southern end. Some of the bands are basaltic and highly amygdaloidal, but these characters have been observed in basic intrusions elsewhere in the area, particularly near contacts with limestone. In thin sections, specimens are felsitic and show intersertal, diabasic, or flow texture. Much of the rock is shattered and schistose. The feldspars are albitized, and sericite, chlorite, quartz, siderite, calcite, and tourmaline have developed. Vesicles are filled with albite (probably end-stage) or combinations of secondary minerals.

Metagabbro porphyry.—Metagabbro porphyry forms two large masses on the north rim of Star Basin, separated by the Star Basin fault zone. The western mass rests with apparent concordance on limestone. The eastern mass is sill-like in form except near the fault zone, where it appears irregular. The masses are lithologically similar but are at different stratigraphic horizons, and, if, as seems probable, they are parts of a single mass separated by faulting and erosion, the connecting portion must have been transgressive. The masses appear to be intrusive, since the upper margin of the eastern mass shows a chilled phase and except for sporadic amygdules lacks structures characteristic of the upper parts of lava flows. Limestone lenses included in the eastern mass have also been recrystallized, and underlying portions of the mass do not suggest lava flows. The actual contacts, however, are concealed by talus.

Jointed, blocky outcrops stained brown by limonite are typical. The rocks are everywhere altered and veined by quartz and coarse calcite. Near the fault zone the eastern mass is shattered and rudely schistose. Allowing for differences in alteration, specimens from the two masses are petrographically similar and range from metadiabase porphyry to coarse metagabbro porphyry. The least-altered specimen obtained from the western mass is a persemic metagabbro porphyry. Plagioclase (Pl. 5, fig. 5) forms partly saussuritized zoned phenocrysts ranging from labradorite to andesine. Actinolite forms abundant sheaves and individual crystals mingled with chlorite, epidote, clinozoisite (?), carbonate, bio-

tite, sphene (?), and secondary magnetite. Some of the resulting aggregates show the octagonal shapes of pyroxene. The groundmass is felsitic and consists essentially of albite and actinolite, with accessory apatite, altered sphene, and magnetite, and minor amounts of other secondary

TABLE 2.—*Mode of metasyenite, by volume*

Component	Per cent
Microperthite	42.1
Saussurite (albite, zoisite, epidote)	19.9
Endstage albite	5.3
Augite	9.6
Chlorite	14.8
Epidote	1.4
Magnetite	3.3
Apatite	2.1
Leucoxene after titanite and ilmenite	1.5
Carbonate	.1
Unknown fibrous mineral	Trace
Total	100.1

minerals. Quartz, present in the groundmass, is for the most part introduced but in part is interstitial and apparently late primary. Pyrite and carbonate, and epidote in part, are introduced minerals. Other specimens examined are more altered. A noteworthy result is complete albitization of plagioclase.

Metasyenite.—Greenish-gray, blocky metasyenite forms a sill traceable from John Brown Canyon to Straight Canyon. The sill is about 700 feet above the base of the limestone and has a maximum thickness of about 100 feet. The rock is everywhere altered and in places is rudely schistose. Its intrusive character is shown by a chilled phase at its upper margin and by baking of overlying shaly limestones to dense hornfels.

The least-altered specimen obtained, from the outlier north of Congress Canyon, is holocrystalline, medium granitoid, hypidiomorphic, interlocking, seriate, massive. The mode as determined by the traverse method is given in Table 2.

Microperthite (Fig. 3) has formed by ragged replacement of orthoclase by albite, which also fills interstices of the orthotectic mineral grains and in this rock is apparently in large part of deuteric origin (Colony, 1923, p. 170-171). Plagioclase is zoned, with saussurite centers and clear albite

rims. Pyroxene (augite?) is partly altered to chlorite. Patches of chlorite, epidote, carbonate, and leucoxene in part may represent biotite. Epidote, chlorite, and carbonate are in part introduced, forming veinlets. Although some of the alteration appears to be deuteric, in large part it is probably the result of hydrothermal activity accompanying mineralization. Evidence of introduction of albite is lacking.

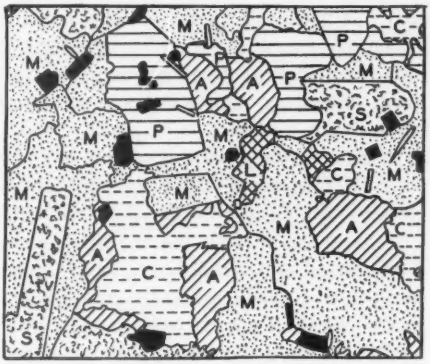

FIGURE 3.—*Metasyenite, north side of Congress Canyon*

Note interstitial habit of much of the albite. A—albite; C—chlorite; M—microperthite; S—saussurite; P—pyroxene; L—leucoxene; black—magnetite; slender prisms—apatite. Line tracing from photomicrograph. × 20.

Granite porphyry.—The principal intrusives in the mapped area consist of granite porphyry,[8] which forms dike-like or irregular masses elongated roughly parallel to the axes of post-Jurassic folding.

The rock is characterized by conspicuous phenocrysts of feldspar or quartz, or both, in an aphanitic or visibly spherulitic groundmass. Unweathered specimens are generally white or light tan but some are gray, blue, or even black. Where sheared, the rock is green, owing to development of sericite. Where weathered, the rock is stained brown by limonite, owing to oxidation of magnetite and disseminated pyrite. Outcrops are rugged and craggy, owing to block jointing.

The granite porphyries (Pl. 5, fig. 6; Pl. 6, figs. 1 and 2) are dopatic rocks with numerous rounded, embayed phenocrysts, 0.4 to 5.0 millimeters in length, of quartz and alkali feldspars in granophyric groundmasses consisting essentially of quartz and orthoclase. The feldspar phenocrysts include perthitic soda orthoclase and soda microcline, "chessboard albite," and normal albite-twinned albite. Many of the feldspar phenocrysts are veined or peripherally corroded by orthoclase (n α less than n α of Na orthoclase; $\gamma - \alpha = .006$; 2V(—) moderate to moderately large). This is interpreted as due to orthotectic reaction. "Chessboard albite," in part at least, has formed by replacement of potash feldspars,

[8] Knopf (1924, p. 35) classified similar masses in the Rochester district as granite porphyry, or alaskite porphyry. Jenney (1935, p. 35-37 and map) mapped a large area of granite porphyry on the west side of the range. Rhyolite porphyry would be a more accurate designation for these rocks in the present area.

for stages in this process have been observed in granite porphyry specimens from the Star Basin, Fourth of July Flat, and the Arizona mine hill. This appears to be a deuteric effect. Later introduction of albite, probably during mineralization, is shown by quartz-albite veinlets in some specimens. In these specimens contemporaneous replacement of feldspar has produced quartz-albite pseudomorphs. Minor constituents are zircon, sphene, apatite, and magnetite. In a single specimen, from the Star Basin granite porphyry mass, scattered pseudomorphs of bleached mica, secondary magnetite, and leucoxene appear to represent biotite phenocrysts.

In general, the granite porphyry has been altered, as have the rhyolitic rocks, but the porphyry is normally less sheared and sericitized. Silicification is the most widespread type of alteration.

Near contacts with the rhyolite formation, the granite porphyry is flow-banded and xenolithic. Hybrid zones are found in which intrusive and intruded rocks are indistinguishable. The principal contact effects in the volcanics appear to have been recrystallization of matrix material to fine-textured quartz-feldspar aggregates. Locally small amounts of biotite and irregular patches of albite have developed. Satisfactory exposures of contacts with limestones have not been found.

Elongation roughly parallel to the fold axes, together with the shearing and accompanying sericitization shown by the intrusions, suggests that the granite porphyry intrusions were emplaced before the close of Jura-Cretaceous (?) orogeny. Emplacement preceded the opening of the ore-bearing fissures, for the granite porphyry contains ore-bearing veins and has been hydrothermally altered. The felsitic and often markedly spherulitic texture of the granite porphyry even in the large masses, and the lack of evidence that important amounts of volatiles were released during crystallization suggests emplacement of highly viscous magma at no great depth.

TERTIARY IGNEOUS ROCKS

Correlation and occurrence.—Widely distributed olivine basalts and other volcanics resting on Truckee (Upper Miocene or lower Pliocene) beds, or on older rocks, have been described from the Humboldt Lake Range by Louderback (1904, p. 301-306, 309-312). Knopf (1924, p. 36-37) and Jenney (1935, p. 47-48) have tentatively assigned olivine basalts and olivine diabase dikes on the west slope of the range to the late Pliocene. In the present area similar olivine basalts, probably of the same age, outcrop along the east base of the range at Cottonwood Canyon and Santa Clara Canyon. At Santa Clara Canyon basalt is associated with trachyte porphyry and trachytic ash, which have not been reported previously from the Star Peak Range.

Santa Clara Canyon area.—The relations at Santa Clara Canyon inferred from scattered outcrops are summarized in Plate 1 and Figure 4. The trachyte porphyry may be an intrusive. The basalts and intercalated trachyte tuff are mapped as a single unit.

The lower basalt is essentially non-porphyritic and normally dense

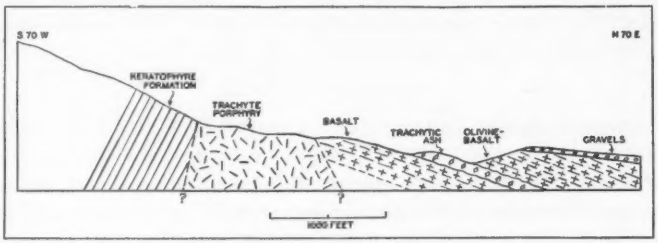

FIGURE 4.—*Section along south side, Santa Clara Canyon*

but in its upper portion it is markedly vesicular. It is gray to black where fresh, brown to red where weathered. Essential minerals, in order of abundance, are labradorite (Ab_4An_6), pyroxene (pigeonite?), and olivine partly altered to iddingsite. Magnetite and apatite are accessories. Carbonate and an unidentified mineral form scattered vesicle fillings. The texture is intersertal or fluidal. The average grain size is about 0.2 millimeter.

The upper basalt differs from the lower chiefly in containing considerable albite, which corrodes the labradorite crystals and fills interstitial spaces.

The intercalated trachytic ash is a hard, light-green, massive rock, consisting of numerous sanidine, oligoclase-andesine, and biotite fragments in a matrix of shards of partly devitrified glass ($n = 1.485 \pm .005$). Felsite and weathered basalt fragments are rare constituents.

In hand specimen the trachyte porphyry is a massive, conspicuously porphyritic felsite, light gray where fresh, stained brown by limonite where weathered. The rock consists of andesine, pyroxene, and biotite phenocrysts in a trachytic groundmass (average diameter 0.15 millimeter) composed of sanidine with accessory disseminated biotite, apatite, magnetite, and a small amount of quartz. Corrosion of andesine by sanidine and resorption of biotite to magnetite granules are apparently primary reaction effects.

Cottonwood Canyon.—A small elliptical knoll rising above the gravels of the Cottonwood Canyon alluvial fan is formed of vesicular olivine

basalt. The top of the knoll slopes gently southeastward, but there is no definite indication of the attitude of the flow.

Diabase dikes.—Olivine diabase dikes, most of which are too small to be shown on the map, are found at a number of places in the area, usually in or near fault zones. The dike rocks are similar to the olivine basalt flows, except for coarser texture and more massive structure. At the head of Star Basin, rapid weathering of narrow dikes in the complex of older rocks has left striking aisles in the cliffs.

QUATERNARY ALLUVIUM

Coarse, subangular gravels cover the canyon floors and form the broad alluvial fans stretching eastward from the base of the range. Similar gravels cover the prominent benches in lower Coyote Canyon and spurs flanking the mouths of Star, Santa Clara, and Tehama canyons. Toward the sides of the canyons the gravels merge into the loose rock debris mantling all but the steepest slopes in the area.

The streams have headwaters in limestone areas and are saturated with calcium carbonate. Evaporation, particularly in the lower parts of the canyons, has caused precipitation of calcite, which in places along present or former channels has firmly cemented the gravels. Conglomerates thus formed can be seen in lower Coyote, Unionville, Big, and Tehama canyons, and in other places, some of which are high in the range. Where entrenchment has taken place ledges of conglomerate outcrop above the present canyon floors.

STRUCTURE

Structure of Star Peak Range.—The complex structure of the Star Peak Range is principally the result of two orogenies of very different types: (1) Jura-Cretaceous (?) folding and faulting; (2) uplift of the range by normal faulting of Pliocene (?) to Recent age. During the Jura-Cretaceous orogeny the Jurassic and older rocks were folded along axes trending generally N. 20°-30° E., oblique to the trend of the range, and in the present area fault zones roughly parallel to the axes of folding were initiated. This deformation was followed by prolonged erosion, by which the Jura-Cretaceous (?) mountains were peneplaned (Jenney, 1935, p. 50). By the second deformation, beginning probably in Pliocene time (Knopf, 1924, p. 36, 41), the Star Peak Range was uplifted as a long, narrow eastward-tilted block, chiefly by displacements along normal faults marking its western margin. Two main stages of uplift have been recognized (Knopf, 1924, p. 41-42; Jenney, 1935, p. 55-57), in the interval between which the range was maturely dissected and igneous activity represented by the basalt flows and associated igneous rocks took place. In the present area faulting

referable to the Pliocene and later uplift can be recognized, but the two stages in the uplift recognized by Knopf and Jenney cannot be clearly distinguished. However, entrenchment of the lower portions of the canyons from Coyote Canyon northward, as shown by gravel-covered benches along the sides of the canyons, may record later stages in uplift.

Structure of present area.—The major structural features of the present area are indicated on Plate 1. South of the area the principal structural feature of the range is a broad, southward-plunging anticline broken by faults. The continuation of this anticline into the present area plunges roughly N. 25° E. from the southern margin of the area nearly to Coyote Canyon. Beyond Coyote Canyon the axis is not clearly traceable, but the apparent plunge is southward. The arch is complicated by faults and minor warps, but its general trend is clearly shown on Plate 1. North of Larned Canyon only the west limb is exposed in the range. From Bloody Canyon northward the dips of the strata on this limb are increasingly steep, and at Tehama Canyon the strata are overturned and dip 80 degrees to the east.

Between Unionville Canyon and Star Basin major fault zones parallel to and west of the main area are associated with folds of roughly parallel trend. The major folds are corrugated by many minor flexures, and owing to these wide variations in strike and dip are found, in places within short distances. Locally, as in Big Canyon, fensters of volcanics have been exposed by erosion of the Star Peak formation from the crests of minor warps.

Faults.—The faults mapped either have actually been observed or have been required by stratigraphic considerations. The maps err on the side of too great simplicity. Many of the faults observed in the mines are not traceable on the surface, owing to lack of key horizons or of exposures. Displacements which appear on the surface as due to single faults are in many cases the sum of displacements on several parallel fault surfaces. Nevertheless, Plate 1 is believed to indicate the principal structural features of the area. The faults range from reverse faults of moderate dip to normal faults with dips as low as 25 degrees. Normal faults are more numerous. Displacements range from a few inches on single fault surfaces to as much as 2000 feet along the major fault zones. The latter figure is the apparent vertical displacement along the axis of the principal anticline in upper Wilson Canyon. The apparent displacements are in most cases parallel to fault dips.

Near faults the rocks have been shattered, dragged, and sheared. Fault surfaces are commonly marked by gouge seams, and in places extensive crush zones occupy the vicinities of major faults. That the strata adjacent

to faults in many places strike parallel to the faults is probably due to drag and rotation. As a result shifts are difficult to estimate for many of the faults.

In the present area faulting of three ages can be recognized: (1) Faulting accompanying the post-Jurassic folding and preceding mineralization; (2) faulting attending mineralization; (3) post-mineral faulting. The major fault zones of the area appear to have resulted from the Jura-Cretaceous (?) orogeny. In the northern part, particularly, the reverse or steep normal attitudes of the zones, or of individual faults within them, the close association of the major fault zones with folds trending parallel to them and oblique to the trend of the range, and the magnitude of the folds suggest origin during Jura-Cretaceous (?) folding. Further evidence given by the structural relations of the ore deposits together with evidence for faulting during mineralization is discussed in a subsequent section.

The major fault zones appear to have been produced prior to mineralization, but in considerable part the displacements along the fault zones took place subsequent to mineralization. The ore deposits show many examples of such post-mineral displacements, which probably represent, in large part at least, the final phase of Jura-Cretaceous (?) orogeny. In part, however, the post-mineral faulting is clearly later and presumably connected with the late Tertiary block faulting. South of the Arizona Mine (Plate 2) a late Pliocene (?) olivine diabase dike is cut off by a narrow fault zone, one of several mapped in the vicinity which trend N. 35° W. to N. 70° W. Displacement along the major fault zones during the late Tertiary is further suggested by the fact that the dike mentioned occupies a fault fissure and by the occurrence of most of the diabase dikes within or near the fault zones. However, this may be due merely to the fact that the zones were lines of weakness, loci favorable to intrusion. The group of roughly parallel dikes in Star Basin, which trend N. 11° W. obviously postdate the development of the fault zone, for one of them cuts the western one of the two faults there shown in Plate 1. The dikes show no effects of shearing.

The possibility that the east base of the Humboldt Range has been determined by Tertiary and later block faulting was recognized by Louderback (1904, p. 342-343). Two features in the present area may indicate such movements. One is what appears to be a low fault scarp traversing the alluvial fans from the mouth of Bloody Canyon to the north side of Coyote Canyon, except where interrupted by channels on the Larned Creek fan. This scarp, however, may be one of the class described by Gilbert (1928, p. 33-39) as "piedmont scarps" and by Longwell (1930, p. 1-13) as "fan scarps", caused by slumping. The second is at the mouth of Santa

Clara Canyon, where the contact of the Tertiary eruptives and the keratophyres may possibly be a fault (Fig. 4).

The schistosity and fracture cleavage shown by the pre-Tertiary rocks have already been mentioned. These features are well developed near the major fault zones. Over most of the area, schistosity, where present, dips west 30 to 60 degrees and generally strikes parallel to the strike of the strata. In lower Tehama Canyon, however, where the strata are overturned to the west, the schistosity dips 60 degrees to the east. Refraction is observed in places where schistosity or fracture cleavage transects lithologic boundaries.

PLEISTOCENE HISTORY

According to Russell (1885, map facing p. 32) the Buena Vista Valley was largely occupied by Lake Lahontan during the high-water stage. On Russell's map the shore line traverses the alluvium near the base of the range. Shoreline features of the lake have not, however, been recognized in the mapped area, although such features are well preserved on the west side of the range.

Obscure indications of Pleistocene glaciation in the region of Star Peak have been reported by King (1878, p. 476) and Blackwelder (1931, p. 913). No special attempt was made during the present investigation to discover such features, but characteristic results of valley glaciation have not been observed either in the region of Star Peak or elsewhere in the area studied.

MINERAL DEPOSITS

GENERAL STATEMENT

Silver, the principal mineral product of the Buena Vista and Star mining districts, has been obtained from quartz bed veins, fissure veins, and stockworks. Stibnite-quartz veins appear to have been second in importance. Gold-quartz veins and other ore bodies, so far as known, have yielded only a small production.

Productive silver-bearing veins and stockworks are found in the lower part of the Star Peak formation and in rhyolite. The tenor of the ores is due in part to the presence of primary freibergite, silver-bearing galena and jamesonite, and pyrargyrite, in varying proportions, and in part to enrichment by supergene argentite and native silver. Enrichment shown by the ores of certain deposits, together with impoverishment in depth, points to supergene enrichment as an important factor in the development of commercial ores.

Antimony-bearing veins have been found in rhyolite and in silty limestone. The principal deposits are in Jackson and Bloody canyons. In these two places stibnite has been found in pods, streaks, and lenses in

quartz veins formed along narrow, steeply-dipping sheeted zones in rhyolite. Ore bodies, so far as known, have been found only in the uppermost flow of the rhyolite formation. Stibnite has been the principal ore mineral, but at Jackson Canyon antimony oxides are reported to have constituted the ore from the upper portion of the productive vein.

The hypogene ore deposits have several significant features in common. Except for differences in types of wall rocks, the deposits exhibit essentially similar wall-rock alteration—chiefly silicification, sericitization, and pyritization. The distribution, the forms, and the internal structures of the deposits indicate a close relationship between mineralization and deformation which took place prior to and during mineralization. In addition, the early stages of mineralization at the different deposits appear to have been essentially similar. It is concluded, therefore, that the deposits were formed during a single epoch of mineralization and under closely similar conditions of moderate temperature and pressure. The relations of the deposits are in accord with the hypothesis, advanced by Knopf and others, that the mineralization was effected by solutions related to the Jura-Cretaceous (?) intrusives.

SILVER DEPOSITS

Location.—The principal silver deposits are the Arizona group south of Unionville and the Sheba-De Soto deposit in Star Canyon (Pl. 1). Silver has also been mined in upper Star Basin, in American Basin, at the Pflum mine just east of Star City, and from other deposits; none of these, however, appears to have yielded any large amount.

Forms.—Three principal forms are shown by the deposits: bed veins, fissure veins, and stockworks. Bed veins in silty limestone have been worked in most of the Arizona group of mines, in the American Basin mine, and the Pflum mine, and have yielded most of the silver. The veins range from 1 inch to 7 feet thick and are essentially conformable to the enclosing strata. All known bed veins are in the lower part of the Star Peak formation.

Fissure-vein and stockwork deposits are found in the lower part of the Star Peak formation and in rhyolite. In the Sheba mine both fissure veins and stockworks connected with them have been explored. Fissure veins or stockworks of minor importance have also been worked in the Santa Clara, Governor Bradley, American Basin, and Champion mines (Pls. 1 and 2).

Bed-vein deposits.—Much of the silver produced in the area was obtained from a remarkable group of bed-vein deposits in the ridges east of Wilson Canyon, mostly in a belt of limestones and associated strata ex-

tending from lower Wilson Canyon to the north slope of Cottonwood Canyon (Pls. 1 and 2). This belt is bounded chiefly by normal faults, along which it is dropped down with respect to the adjacent rocks. In the area covered by Plate 2 the thickness of the Star Peak beds is nowhere more than a few hundred feet, and in places erosion has exposed the underlying volcanics. The strata have a general easterly dip, but dips and strikes vary widely owing to minor folds, and trends of contacts are more reliable as indicators of formational attitudes. There are numerous faults, of which those traceable on the surface are only a small proportion of those seen in the mines. The majority are normal faults trending N. 10° E. to N. 50° W. and dipping 25° W. to 60° E.

The bed veins are found at approximately the same stratigraphic horizon, in silty, laminated, or shaly limestones approximately 40 to 50 feet above the top of the basal clastics. This correspondence in position, however, does not necessarily indicate that the deposits represent a single, originally continuous bed vein as suggested by Ransome (1909, p. 40). Bed veins are found at several horizons in the Arizona mine within a vertical distance of about 25 feet. Furthermore, evidence of pre-mineral faulting makes it doubtful that the strata were continuous at the time of mineralization. It seems more probable that the correspondence in position is due to formation of channels favorable to bed-vein development only in certain silty limestone strata which yielded during deformation by separation along closely spaced lamination surfaces.

The largest and most productive deposit of this group was explored by the workings of the Arizona mine. This deposit was examined and described briefly by Ransome (1909, p. 38-40) and others, who recognized its remarkable nature. The deposit, however, merits more detailed description. It is the largest and best-exposed example of the bed-vein type within the area. Furthermore, the structure of the deposit is important in connection with the problem of supergene enrichment of the silver deposits and furnishes significant information with respect to the influence of deformation on mineralization.

The mine was only partly accessible at the time of visit, and only an incomplete plan of the workings was available (Fig. 5). In the inaccessible Manitowoc or northeastern workings, a bed vein striking N. 30° W. (J. D. Hague, 1870, p. 311) and dipping gently southwest was worked. In most of the accessible parts of the Arizona mine proper, which extends westward and southward from the Stewart fault zone, a single eastward-dipping bed vein was worked.

In the Arizona workings the strata in general dip eastward and strike N. 5°-40° W. Strata and bed vein are displaced a few inches to 15 feet by normal faults striking N. 20°- 50° W. (with few exceptions) and dipping

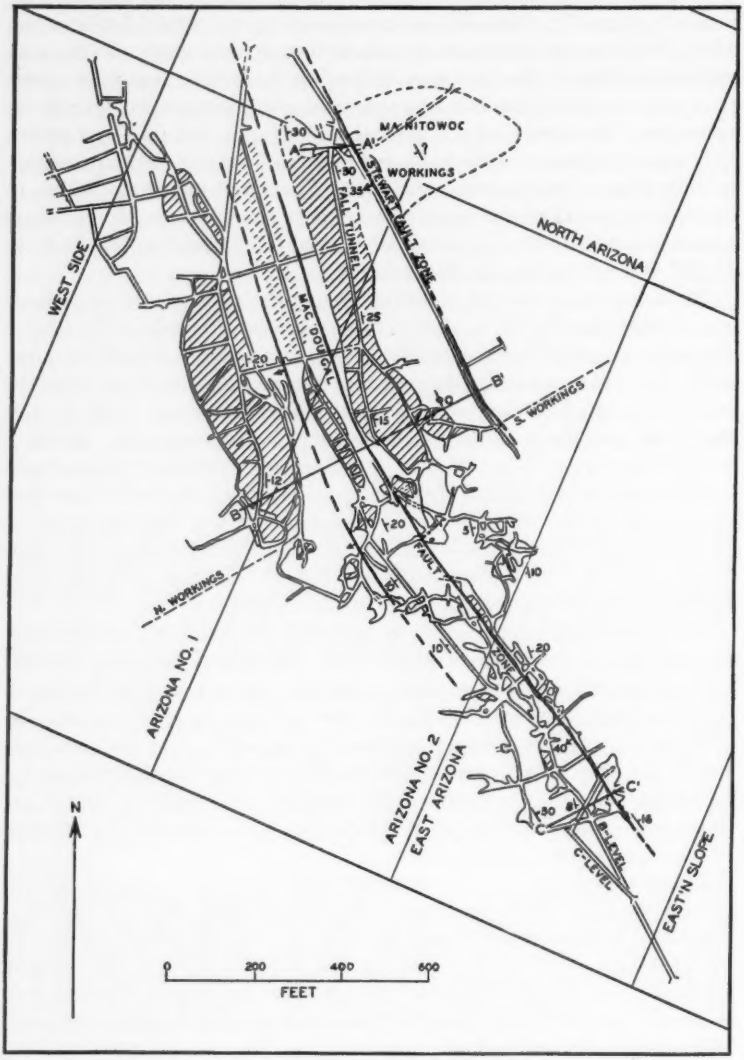

FIGURE 5.—*Plan of Arizona mine*

A-level not shown. Workings in the northwestern part of the mine not completely represented. Lined areas are filled stopes.

25°-50° W. A few apparent reverse faults are probably rotational. Eastward-dipping normal faults are rare. Where the faults are closely spaced narrow zones of relatively large displacement occur. These zones trend N. 20°-40° W. and dip 30°-40° W. Owing to the divergence of fault-zone and limestone trends, the northern part of the deposit (Fig. 6) is divided

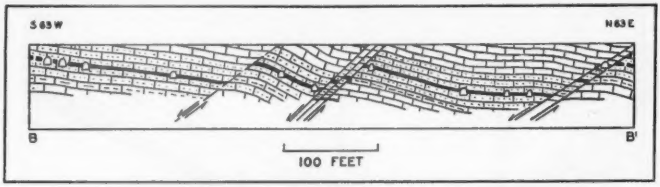

FIGURE 6.—*Cross section of the Arizona deposit*
Along BB' of Figure 5. Bed vein solid black.

into parallel "troughs" plunging gently south-southwest. The east wall of each is a westward-dipping fault zone, and the west wall is a fault segment of limestone.

Near faults the limestone and vein material have been crushed and flexed, and the vein in places has been pinched out. Fault surfaces are marked by seams of black carbonaceous gouge containing fragments of brecciated vein material. Thick gouge zones are found along the Stewart and MacDougal fault zones.

The main bed vein ranges from a few inches to 7 feet thick, averaging about 2.5 feet. It appears to have followed essentially a single stratigraphic horizon throughout the mine. The vein contacts are frozen except near faults. The vein walls, usually sharp and parallel, clearly indicate separation of the silty limestone along lamination surfaces. Where a contact crosses from one lamination plane to another the resulting "jump" is usually angular. Many of the limestone partings in the vein are likewise angular. Offshoots of the vein along fractures show similar relations at many places. These characteristics were early recognized, and Raymond (1873a, p. 205) remarked that the vein followed the stratification but resembled a fissure vein. Ransome (1909, p. 40) concluded:

"There is no evidence of any important replacement, and the quartz has crystallized in open spaces produced by the separation of the limestone along one or more bedding planes. The banded structure (of the vein) suggests that the separation was effected by successive movements, each followed by the deposition of quartz."

In general the writer's observations support the conception of deposition in open spaces resulting chiefly from separation of silty limestone along closely spaced lamination planes. Cavities and comb structure in

the vein quartz are further indicative of open spaces during early quartz deposition. In addition, thin sections of the vein material yield ample evidence of successive movements, during mineralization, which sheeted early quartz parallel to the vein surfaces and opened channels for the

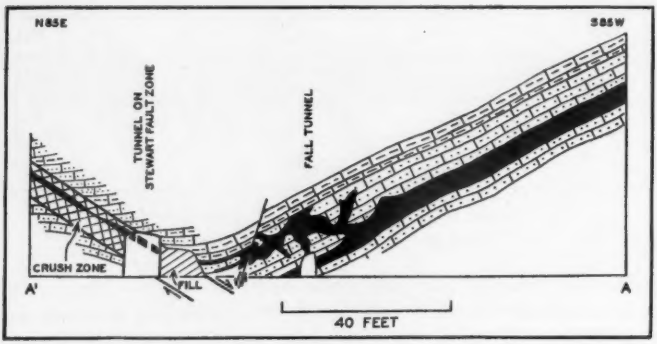

FIGURE 7.—*Structure section, Arizona deposit*
Along line AA' of Figure 5. Minor details omitted. Vein matter solid black.

entrance of later vein minerals. It seems unlikely that over any large part of so great an area as that occupied by the Arizona vein open spaces along bedding planes could exist simultaneously in strata of gentle to moderate dip, beneath the weight of the 5000 to 7000 feet of overlying strata inferable during mineralization. It seems possible, however, that over a considerable period of time successive shearing movements parallel to bedding planes might provide open spaces at a structurally favorable horizon first at one point and then another, with the result that a vein of considerable area might ultimately be formed. A mechanism of this sort has been proposed by Johnston (1938, p. 23-34), in explanation of fissure veins at Nevada City, California.

The Arizona bed vein, however, was not formed solely by filling of open spaces. Metasomatic replacement was an essential phase of deposition of the metallic hypogene minerals. Replacement of inclusions and partings in bed vein ore by quartz has been an important cause of the banding in the ore. Furthermore, local replacement is indicated by marginal relations of the type shown just west of the Stewart fault zone in Figure 7.

Smaller bed veins, two of which are shown in the cross sections (Figs. 7 and 8), are found at several horizons both above and below the main vein.

Although a remarkably persistent bed-vein form characterizes the deposit, there are two noteworthy exceptions. According to Raymond (1872, p. 135) ore was found along the Stewart fault zone for 400 feet south from the intersection with the Fall tunnel. The nature of this

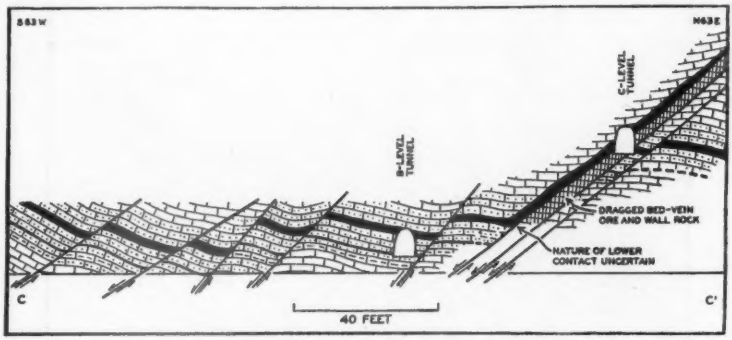

FIGURE 8.—*Structure section, Arizona deposit*

Along line CC' of Figure 5. Section west of the B-level tunnel is generalized. Vein matter solid black.

"ledge" was not, however, clearly understood. A lawsuit arose in 1870 from uncertainty as to whether this ledge constituted (1) a separate vein, (2) a spur of the Arizona bed vein connecting upward with the western extension of the Manitowoc bed vein, which apparently is at a higher level than the Arizona main bed vein, or (3) vein matter dragged along a fault zone offsetting two parts of the same vein. At the north end of the mine (Fig. 7) the "ledge" is a definite vein 1-12 inches thick along a fault fissure within the Stewart fault zone. The relations of the vein to the Arizona and Manitowoc veins could not be determined, owing to deterioration of the workings, removal of much of the ledge, and post-mineral deformation.

A similar fissure vein stoped at intervals for about 400 feet along the MacDougal fault zone can be traced along the line of Fig. 8 downward and westward into the main bed vein. The fissure vein apparently constitutes a spur of the main bed vein along a fissure within the MacDougal fault zone, but the relation of this vein to the bed vein east of the fault zone could not be determined. From the relations seen at C it seems probable that the Manitowoc vein is the upfaulted eastern extension of the Arizona vein, of which the fissure vein along the Stewart fault zone is a spur. At neither place was any indication seen that the fissure veins continue into depth.

The principal displacements on both the MacDougal and Stewart

fault zones were clearly post-mineral, for the bed veins are offset by them, and the fissure veins in places pinch out and are replaced by seams of carbonaceous gouge. However, the formation of veins along fissures parallel to the post-mineral fault surfaces would seem to indicate that

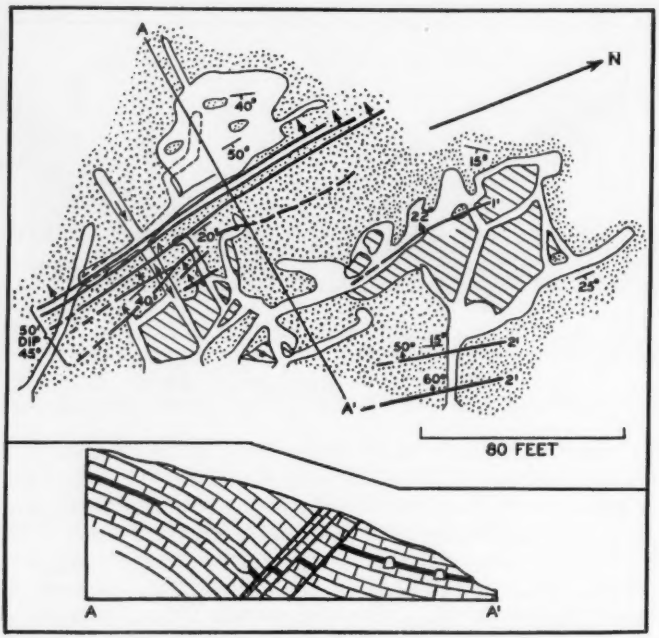

FIGURE 9.—*Plan and section of the Wheeler mine*

Based on pace-compass survey. Vein solid black in section. On map, heavy lines are faults, figures at terminations indicate dip-slips in feet.

deformation along these zones began prior to mineralization. Deformation during mineralization is indicated by distribution of sulphides in the fissure veins in streaks and bands formed by replacement of quartz along sheet fractures parallel to the walls. Initiation of faulting in the area of Plate 2 prior to mineralization is further indicated by the occurrence of the Governor Bradley vein along a fault (H, Pl. 2).

The deposits explored by the Wheeler mine (Fig. 9), and the Inskip, North Star, Millionaire, Lucky Dog, and Cottonwood mines are bed veins similar to the main bed vein of the Arizona deposit. In the northern part of the area (Pl. 1), similar bed veins are found in the American Basin and Pflum mines.

Fissure veins and stockworks.—For purposes of discussion of the mineral deposits of the area the distinction of fissure veins and stockworks is useful, since it serves to designate two of the principal forms exhibited

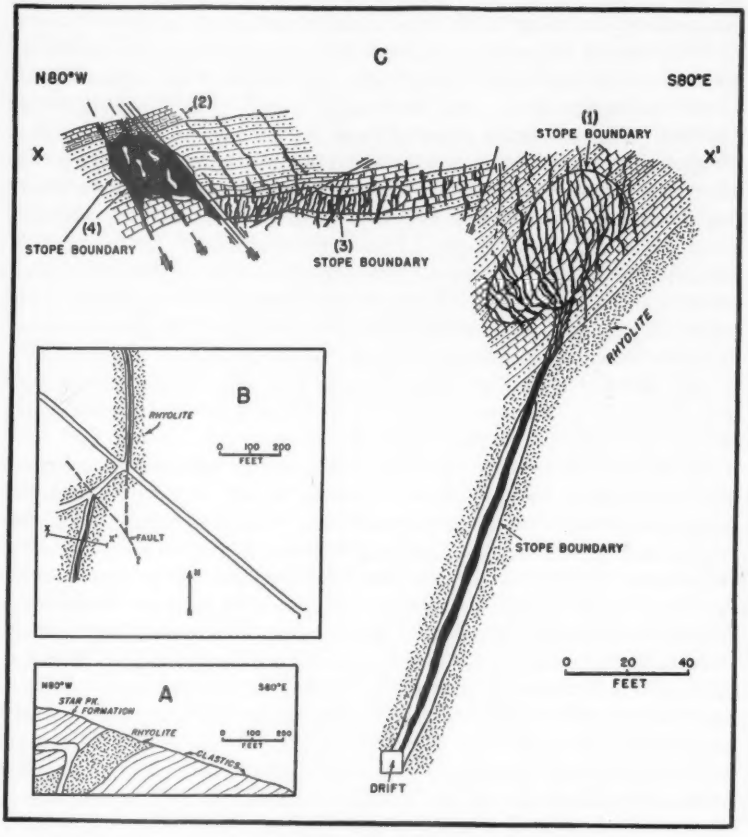

FIGURE 10.—*Sheba deposit*

A—diagrammatic section showing relation of C to the structure of the vicinity. B—plan of part of workings at level of drift at base of C; line XX′ is the line of section C. C—section across Sheba deposit, showing types of ore bodies. Vein matter solid black.

or approximated by deposits of the area other than bed veins. However, the fissure veins and stockworks are closely related to one another. The features of the Sheba deposit serve to illustrate this point and also to exemplify distinctive features of each of the two forms.

At the time of visit, the workings of the Sheba mine were only partly accessible. However, Figure 10 shows the principal types of ore bodies encountered in the deposit and the relation of the deposit to the enclosing rocks. Judging from descriptions by Raymond (1873a, p. 208-211; 1875, p. 264-265) and Ransome (1909, p. 42-43) these ore bodies, with their inaccessible extensions, constituted a major portion of the deposit.

The principal ore bodies were found in the lower 100 feet of interbedded limestones and tuffaceous basal clastics of the Star Peak formation, between dark-gray, shaly carbonaceous limestones (Fig. 10-C -(2)) and rhyolite. The ore shoots extended from the surface on the side of Star Canyon northward into the hill for about 300 feet (Ransome, 1909, p. 42) along the strike of the strata. The eastern part of the deposit is connected with a fissure vein 1 to 3 feet thick in the underlying rhyolite[4] south of the adit tunnel shown. This vein strikes N. 10° E. to N. 15° W. and dips 70° to 80° W. In some places the vein contacts are frozen; elsewhere smooth, gently curving, gouge-coated walls enclose the vein. In some parts of the vein the full width is ore; in others elongate inclusions, large or small, mark branches in the controlling sheet structure. For about 250 feet along the strike the vein has been stoped upward from the drift shown. A similar vein north of the adit tunnel is probably the same vein offset along a fault (Fig. 10-B).

At the contact between rhyolite and overlying tuffaceous sandstones the vein changes abruptly to a stockwork 4 feet wide which broadens about 15 feet above into a wide stockwork consisting chiefly of veinlets along fracture cleavage surfaces dipping steeply east. This whole mass was stoped (Fig. 10-C) for more than 150 feet along the strike. A tunnel leading from the top of the chamber connects with workings to the west. The strata here are traversed by numerous veinlets, the most prominent of which are formed along steeply dipping fracture cleavage. Particularly in thin limestone beds fracture cleavage is closely spaced, and veinlets are correspondingly numerous. The outline of an extensive stope on one such bed is shown in Fig. 10-C at (2). At (4) is shown the cross section of a stope approximately 150 feet long, 30 feet high, and 20 feet wide which trends N. 10° E., parallel to the strata, and is confined to massive beds of tuffaceous sandstone. This chamber appears to occupy the intersection of the sandstone and a series of eastward-dipping premineral normal faults and to have resulted from partial replacement of adjacent fractured rock by vein matter which entered chiefly along the

[4] This rhyolite resembles granite porphyry in containing abundant phenocrysts of quartz and alkali feldspars, but its regular contacts with the overlying clastics and the absence of contact effects in specimens of adjacent rocks have led to the conclusion that it is merely an unusually porphyritic rhyolite. The groundmass texture has been obscured by shearing and hydrothermal alteration (Pl. 6, fig. 3).

fault fissures. The form of the ore body inferred is shown in the section.

From Ransome's (1909, p. 42) description the main fissure vein would appear to fill the channel along which the vein material in the overlying sediments was supplied. This is true of the stockwork at the upward end of the fissure vein, but the ore bodies to the west are connected with other channels. Such channels appear to be represented by two zones of fissure veining exposed in the main tunnel west of the fissure vein. One, striking about N. 40° E. and dipping 60° W., is in rhyolite; the other, striking N. 35° E. and dipping 65° E., is in limestone.

It is evident that the deposit as a whole is a mineralized zone, in which forms of ore bodies differ from place to place in correspondence to the patterns of channels produced in different types of rock by fracturing. Marked refraction of fissures passing from rhyolite into the sediments is a noteworthy feature of the deposit.

The De Soto mine was inaccessible at the time of the writer's visit, but according to Ransome (1909, p. 43) the deposit is the southern extension of the Sheba deposit and is similar in form.

The stockwork type of deposit, as an independent form, is best exemplified by the deposit at the Champion mine, in Star Basin (C, Pl. 1). Mineralization of the brecciated rocks on either side of the fault zone here has resulted in a wide quartz replacement stockwork about 1500 feet long and 400 feet wide. The deposit consists of a set of major westward-dipping fissure veins connected by smaller veins along several sets of cross fractures. The principal veins strike N. 10° E., approximately parallel to the fault zone, and have clearly resulted by filling and replacement along fractures opened during the deformation which initiated the fault zone.

The close relationship between forms of deposits and structural features produced by deformation in different types of wall rocks is further indicated by features of a group of deposits along the west side of American Basin. Here a mineralized zone in the lower 150 feet of limestones and interbedded basal clastics of the Star Peak formation is exposed by a series of trenches and adits. The zone resembles the Sheba-De Soto deposit, consisting of veins along fault fissures of small displacement striking N. 55° W. and dipping 75°-90° N. and connected with stockworks in certain beds. An adit at the southern end (G, Pl. 1)—the only one open—gives access to the mine shown in Fig. 11. The deposit exposed is of interest because it shows points of resemblance to both the Sheba and Arizona deposits.

From the portal the adit follows a fissure vein 4 to 12 inches thick along a fault fissure in silicified, sericitized, pyritized, talcose and tuffaceous sandstones. At 160 feet the fissure vein intersects the overlying

silty limestones and abruptly changes to a stockwork. In the inclined stope west of this point the fissure vein is absent, and the vein material is largely confined to a single bed of silty limestone, in which it forms a stockwork or a typical bed vein. The actual transition from fissure vein to bed vein could not be seen, but there seems no reason to doubt their

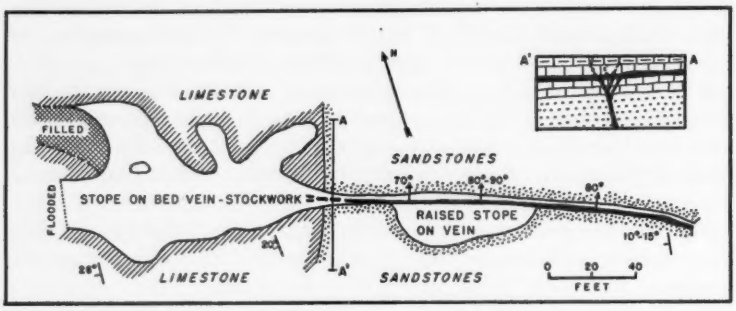

FIGURE 11.—*Plan and section, American Basin mine*
Based on pace-compass traverse. Vein matter solid black.

continuity. The fissure vein is similar to small fissure veins seen in the Sheba mine. The bed vein, however, is similar to the Arizona bed vein both in form and in appearance of ore.

The fissure-vein type of deposit in its simplest form is exhibited at a prospect in Santa Clara Canyon (Pl. 1, B). The vein exposed here is a fissure vein 6 to 12 inches wide, formed by filling and replacement along a narrow sheeted zone in rhyolite. It is noteworthy that the vein and sheeted zone, which dip 70° SE., strike N. 30° E., roughly parallel to the major fault zone which traverses Santa Clara Canyon below the prospect.

Bed-vein ore.—Bed-vein ores are predominantly composed of gray to milky quartz of variable but prevailingly coarse texture. Scattered cavities lined with quartz crystals are present, and in some specimens comb structure is apparent. Banding parallel to the walls is usually conspicuous. It is due to (1) thin carbonaceous seams and silty limestone partings; (2) streaks of dark, fine-textured quartz representing silicified partings whose original angular outlines are indicated by the distribution of unreplaced particles of carbon and other impurities; (3) ribboning due to shearing parallel to the vein surfaces; and (4) dissemination of metallic minerals in one or more layers parallel to the vein surfaces.

In much of the ore the metallic minerals form minute particles disseminated through bands of medium- to fine-textured quartz. The soft

shadowy gray of such bands is characteristic. Pyrite is the only mineral visible in some specimens, but in others tetrahedrite, galena, sphalerite, or other sulphides can be detected. In places, however, irregularly distributed streaks and bunches of coarser-textured sulphides are found. Ore high in silver contains greenish-black sooty spots or masses, the

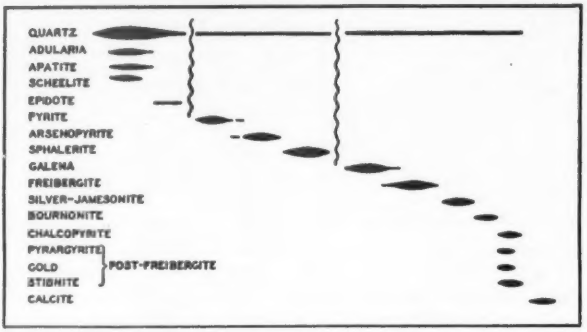

FIGURE 12.—*Sequence of mineral development in the silver ores*
Periods of deformation are indicated by vertical wavy lines.

essential constituents of which are argentite, covellite, chalcocite, native silver, and cerussite. Limonite, malachite, azurite, and copper sulphates stain the oxidized ores.

Ore of fissure veins and stockworks.—Fissure-vein and stockwork ores consist principally of coarse milky quartz, massive or with rude comb structure, containing scattered quartz-lined cavities. Ribbon structure is locally present. Generally the metallic minerals are sparsely distributed in streaks and bunches, but in places are disseminated in bands. They are usually coarser in texture than corresponding minerals of the bed veins. Oxidized ores are stained by limonite and frequently by copper carbonates and sulphates. The fissure veins contain numerous wall-rock inclusions, many of which show partial replacement.

Hypogene mineralization.—A composite summary of the hypogene mineralization recorded in various silver ores is given in Figure 12. The principal hypogene sulphides are pyrite, arsenopyrite, sphalerite, freibergite, and silver-bearing galena and jamesonite; the last three are the most abundant hypogene silver minerals. Analyses of freibergite and silver-jamesonite by Burton (1868, p. 36-38) are shown in Table 3.

An assay in duplicate of freibergite from the Sheba mine made by Mr. Richard Forbes indicates, after allowance for 2.7% of impurities, a silver

content of between 11.3% and 11.5%. Whether the higher percentage of silver shown by Burton's analysis is due to variation in the composition of freibergite or to microscopic impurities not detected by Burton is not known. Sufficient jamesonite for assay could not be isolated by the writer,

TABLE 3.—*Analyses of silver-bearing sulphides*

	I Per cent	II Per cent
S	19.06	24.35
Sb	29.26	27.35
Pb	43.86	.....
Ag	6.14	14.59
Cu	1.55	27.40
Fe	.05	4.27
Zn	.....	2.31
Insol	(1.53)	0.35
	99.92	100.62

I. Mean of two analyses of jamesonite from Sheba mine. Insoluble residue deducted.
II. Analysis of freibergite from De Soto mine.

but microchemical tests indicate the presence of silver in jamesonite which shows no trace of inclusions (Schneiderhöhn and Ramdohr, 1931, p. 402). Silver shown by microchemical tests to be present in galena is probably for the most part contained in an unidentified mineral [5] (Pl. 6, fig. 4) forming numerous microscopic inclusions, as well as small but widely distributed particles in other ores.

Pyrargyrite has been reported to form by both hypogene and supergene processes (Emmons, 1933, p. 413). It is a late mineral in the present ores, but its textural relations are identical with those of known hypogene minerals, and its distribution is unrelated to the distribution of known supergene minerals. Hypogene origin is thus indicated. Gold has replaced freibergite in ore from the Sheba mine, and freibergite and possibly galena in ore from the Santa Clara deposit. The ores of the Governor Bradley and Cottonwood mines (Pl. 2) have been reported to contain gold, but the mineral is not present in specimens examined microscopically by the writer. Stephanite (possibly supergene (Emmons, 1933, p. 412)) reported from

[5] White, slightly greenish against galena, bladed or acicular, slightly softer than galena, strongly anisotropic. Apparently (microchem.) a silver-lead sulphantimonide. Resembles jamesonite, but does not etch with HNO_3. Possibly silver-bearing jamesonite. (*Cf.* Schneiderhöhn and Ramdohr, 1931, p. 402-403.) Not recognized in ore specimens containing identifiable silver-jamesonite.

the Sheba mine (A. Hague, 1877, p. 731) has not been found by the writer. Bournonite is abundant in ore from the Santa Clara mine but has not been identified in other silver ores. Stibnite is a minor constituent in ore from the Sheba mine.

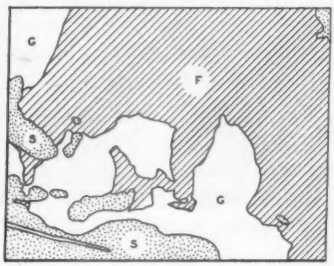

FIGURE 13.—*Characteristic intergrain boundaries in hypogene ore*

G—galena; F—freibergite; S—sphalerite; Q—quartz. Arizona mine. Line tracing from photomicrograph. × 42.

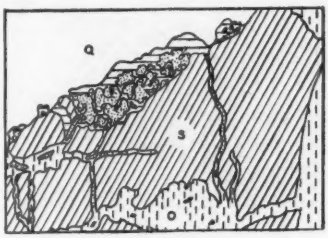

FIGURE 14.—*Selective replacement of freibergite (stippled) by supergene argentite (horizontal ruling)*

Sphalerite (S) not attacked. O—oxidation products, chiefly cerussite, with spots of native silver (black). Q—quartz. American Basin mine, line tracing from photomicrograph. × 119.

Hypogene gangue minerals other than quartz are only locally abundant. Adularia is present in small amounts except in ore from the Sheba deposit. Scattered grains of apatite are present in bed-vein ores. Small grains of scheelite, associated with epidote, have been found in a few specimens of ore from the Arizona bed vein. About 1919, according to local report, several hundred pounds of high-grade scheelite ore were removed from an unknown place in the Arizona mine. Specimens of this material examined by the writer consist of coarse, honey-colored scheelite, intergrown with coarse quartz. Barite in cavities in ore of the American Basin deposit may be supergene. Calcite, locally forming coarse-textured masses, appears in part to be late hypogene. In part it is clearly supergene, for it has replaced quartz along open fractures containing other supergene minerals. Massive calcite from parts of the Arizona and Wheeler deposits is fluorescent, probably due to a small amount of manganese indicated by qualitative chemical tests. Sericite in the veins is for the most part an alteration product of feldspathic wall-rock inclusions. Ore from the Champion mine and a few specimens from the Sheba mine exude free sulphur along fractures when heated above 120° C.

The hypogene mineralization was accomplished in successive stages, delimited by periods of fracturing. During each stage deposition took place by metasomatic replacement of earlier minerals along shear zones and fractures (Pl. 6, figs. 5 and 6). Quartz was deposited during all

stages. Characteristic textures of the hypogene ore minerals are illustrated by Figure 13. Successive metasomatic replacements have resulted in intergrowths of varied textures. An important feature, however, is the characteristic clean-cut nature of intergrain boundaries.

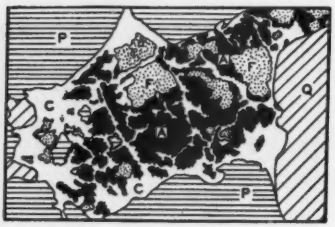

FIGURE 15.—*Selective replacement by supergene sulphides*

Pyrite (P) partly replaced by freibergite (F), in turn replaced by sooty supergene argentite (A) and later intergrown chalcocite and covellite (C). Pyrite not replaced by minerals other than freibergite. Q—quartz. Arizona mine. Tracing from photomicrograph. × 110.

FIGURE 16.—*Textures of supergene sulphides*

Freibergite (F) partly replaced by a sooty intergrowth of argentite and covellite (AC); Q—quartz. Arizona mine. Tracing from photomicrograph. × 15.

Supergene minerals of Arizona mine.—The paragenesis and development of supergene minerals in the silver ores is shown most clearly in ore from the main bed vein of the Arizona mine. The known hypogene metallic minerals in this ore are pyrite, arsenopyrite, sphalerite, galena, freibergite, and silver-jamesonite(?). The principal silver minerals appear to have been argentite and native silver.[6] The following features, disclosed by microscopic examination of the ore, relate to the origin of argentite:

(1) Argentite formed later than freibergite. The relation of argentite to silver-jamesonite(?) has not been observed.

(2) Argentite has formed in large part by replacement of freibergite (Fig. 14), in small part by replacement of sphalerite and galena.

(3) In all specimens examined argentite is associated with, and frequently intergrown with, covellite, and in places chalcocite. In large part covellite and chalcocite are later than argentite (Fig. 15).

(4) In rich ore, sooty greenish-black spots or masses partly fill cavities left by removal of known hypogene sulphides. Ore of this type has been reported to assay up to 1750 ounces of silver per ton and 0.5 ounces of gold (Ransome, 1909, p. 40). These masses (Fig. 16) are mixtures of argentite, covellite, chalcocite, and oxidation products replacing freiberg-

[6] According to Raymond (1869, p. 123) the ore close to the surface, however, was largely silver chloride.

ite and associated sulphides along intricate networks of irregular fractures. The resulting textures are in sharp contrast to textures of known hypogene minerals but are similar to textures resulting from replacement of the sulphides by oxidation products (Fig. 17). Such textural features are in general shown by argentite, covellite, and chalcocite even when megascopically invisible.

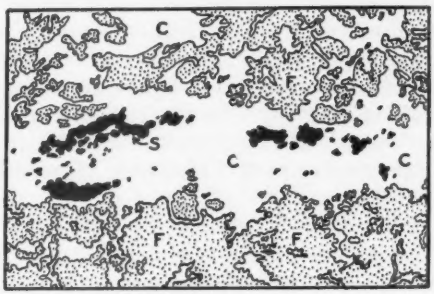

FIGURE 17.—*Textures of partly oxidized ore*

Freibergite (F) partly replaced by cerussite (C) containing native silver (S, solid black). Arizona mine. Tracing from photomicrograph. × 85.

(5) In places the mixtures are related to fractures in quartz which, unlike fractures produced during hypogene mineralization, have not been healed by later quartz and which therefore appear to be post-hypogene. The contrast in texture and textural relationships between covellite-chalcocite-argentite mixtures and intergrowths of known hypogene sulphides, the textural similarities between the mixtures and known oxidation products, the sooty character of the visible mixtures, the leached condition of the ore in which the mixtures occur, their late position in the sulphide sequence, and their occurrence along fractures which appear to be post-hypogene are evidence of their supergene origin. The formation of argentite, therefore, represents a supergene enrichment of the bed-vein ore.

Covellite is the most abundant supergene sulphide. Covellite and chalcocite have partly replaced other sulphides, but apparently not pyrite and arsenopyrite (Fig. 15).

Native silver, wherever observed, forms small particles associated with limonite, cerussite, malachite, azurite, and copper sulphates, and in considerable part is disseminated in cerussite (Fig. 17). In part it is a direct oxidation product of silver-bearing sulphides; in part it is clearly introduced and has resulted in enrichment of the ore in silver. Cerargyrite has not been detected by the writer. Gold is present in oxidized ore from the Manitowoc mine (U. S. N. M. Spec. 33708).

Supergene minerals of other silver deposits.—The above description is applicable in general to other silver deposits of the area, ores of which have been examined microscopically. Supergene argentite is present in ores from the Wheeler, Inskip, and American Basin mines; native

silver in ores from the Wheeler, Governor Bradley, American Basin, and Champion mines. Supergene covellite is present in ores from all silver deposits and is accompanied by supergene chalcocite in ores from the Inskip, Sheba, and American Basin mines. Although evidence of enrichment of ores by supergene silver minerals has not been observed in ores of all the deposits, the presence of one or more supergene sulphides in ores from every deposit indicates that the deposits as a group have been affected by similar processes of supergene sulphide formation. The possibility cannot be overlooked, moreover, that although a considerable number of polished specimens were examined microscopically, the element of chance involved in selection of specimens from comparatively lean ores has caused supergene sulphides to be overlooked in certain deposits.

Limonite and cerussite are the most abundant oxidation products. Limonite stains quartz of the outcrops, fills cavities left by leaching of sulphides, and stains ore beneath the surface. Manganese oxides, accompanied in places by copper sulphates and carbonates, are present in small amounts in most of the oxidized ores. Cerargyrite, reported from several deposits, has not been detected by the writer.

SUPERGENE ENRICHMENT OF SILVER DEPOSITS

Importance and extent.—Evaluation of the importance of supergene enrichment in silver deposits from which the rich ore bodies have long since been largely removed, and concerning which only fragmentary historical data are available, is inherently difficult. Nevertheless, it is believed that at least in most of the deposits supergene enrichment of lean protore has been the factor delimiting profitable from unprofitable ore.

There are two reasons for this conclusion: (1) The ores of certain deposits show clear evidence of enrichment by supergene argentite and native silver, and (2) in those deposits for which information is available, the rich ores have been found near the surface. Followed into depth the ores were not workable, a circumstance for which, in most instances, no explanation other than secondary enrichment has been found.

Decline in the value of the ores with depth is most clearly illustrated by the Arizona mine. Development of the mine proceeded in general southward from the north end of the deposit and therefore followed the plunge of the structural troughs constituting the deposit. The values of the ores mined in successive years, therefore, give a rough measure of changes in the tenor of the ore as the deposit was followed into depth. In 1869 (J. D. Hague, 1870, p. 311) the ore mined averaged about $100 per ton. In 1875 the average (pulp assay) was $30 per

ton (Raymond, 1877, p. 186), and in 1908 (Ransome, 1909, p. 40) the average for all faces of the vein was about $21 per ton. The decline in the tenor of the ore as the deposit was mined down the plunge of the structure corresponds to conditions expectable if the silver content is largely dependent on supergene enrichment. Though the greatest enrichment appears to have taken place at the north end of the deposit, supergene sulphides, in some specimens accompanied by native silver, are present in samples collected near the south end of the south workings.

The other deposits in general appear to have had a similar history. At the Wheeler mine, the richest ore was found near the surface (Raymond, 1877, p. 186). In the Pflum mine, according to the owner, ore high in silver was also found near the surface. At the Champion mine rich ore found near the surface changed within 75 feet downward into a lean primary sulphide ore, and the mine was abandoned. At the Governor Bradley mine ore assaying $63 to $2765 per ton in silver and $20 to $180 per ton in gold was found near the surface (Whitehill, 1873, p. 58), but the deposits evidently failed to be workable in depth. A similar history is indicated for the Cottonwood, Millionaire, and Santa Clara deposits.

A possible exception to the general rule is the Sheba-De Soto deposit. The principal sulphides in ore from this deposit are pyrite, sphalerite, freibergite, and silver-bearing galena and jamesonite. Arsenopyrite, chalcopyrite, pyrargyrite, stibnite, and gold are minor constituents. The ore is partly oxidized. Polished surfaces and thin sections show supergene covellite and chalcocite, together with malachite, cerussite, and anglesite. Whether the specimens examined are representative of commercial ore is uncertain, for only vestiges of the original ore remain in the mine, and the material on the dumps is largely the residue of concentration. The importance of supergene enrichment is correspondingly difficult to evaluate.

From descriptions by Stretch (1867, p. 51) and Raymond (1873a, p. 211) it would appear that hypogene sulphides, particularly freibergite, constituted the bulk of the ore mined from 1866 to 1875, but whether these descriptions are applicable to the bonanzas mined earlier is uncertain. The only probably supergene silver mineral reported (Stretch, 1867) is native silver. The extent to which supergene silver minerals unrecognizable at that time were present is also uncertain, but the fact that neither covellite nor chalcocite was reported suggests that supergene silver minerals may have been more abundant than supposed. However, supergene silver minerals have not been detected in polished specimens.

That the deposit was apparently not workable in depth is not necessarily indicative of significant enrichment. The dependence of the vein material on channels produced by fracturing suggests that the stratified

rocks were structurally more favorable than the rhyolite to the development of workable ore bodies. Lateral spreading of the mineralizing solutions due to ponding beneath the shaly limestones (Fig. 10B-(2)) forming the "hanging wall" of the deposit is suggested by the apparent restriction of the ore bodies to the strata beneath and by the fact that the fault-fissure veins of the westernmost stope split and dwindle on passing upward into the shaly limestones.

Development of supergene sulphides.—It is generally accepted that the principal active agents in the oxidation of pyritic sulphide deposits are dilute solutions of iron sulphates and sulphuric acid derived by oxidation of pyrite. Sulphate ion in the silver-mine waters of the present area is indicated by crusts and innumerable minute crystals of gypsum coating fractures in vein material or in gouge seams in deposits in limestone. Damp ore from the Arizona and Wheeler mines, from the latter particularly, becomes coated with a thick powdery efflorescence of epsomite and tamarugite(?) on drying. The magnesium is probably derived from dolomite in the lower limestone strata.

With regard to development of supergene sulphides in the silver deposits of the present area the important substances present in the downward-percolating meteoric solutions resulting from oxidation of the upper portions of the deposits were presumably ferric, cupric, and silver sulphates. In silver ores of the Rochester district Knopf (1924, p. 47) found supergene argentite formed for the most part by replacement of sphalerite. In ores of the present area argentite has formed mostly by replacement of freibergite (Fig. 14). Pyrite and arsenopyrite appear to have been immune to replacement by supergene sulphides (Fig. 15). The immunity of pyrite and the preponderance of covellite over chalcocite suggest that the concentrations of ferric and cupric sulphates were relatively high, and of ferrous and cuprous sulphates relatively low, in the enriching solutions (Zies, 1916, p. 429, 451-452).

Depth and time of enrichment.—The depth to which sulphide enrichment has affected the silver ores is unknown. As far as known none of the deposits has been explored to depths greater than about 300 feet. Part of the enriched zone has undoubtedly been removed by erosion, for supergene sulphides are found at the vein outcrops. Enrichment is clearly unrelated to the present water level, since at least in certain mines supergene sulphides are found above the water table.

In the Rochester district the supergene sulphide zone has a minimum vertical range of 800 feet, with an unknown amount removed by erosion of the upper part of the zone (Knopf, 1924, p. 50). The water table at present is below the downward limit of ore sufficiently enriched by

argentite to be minable. Knopf has tentatively referred sulphide enrichment to the period succeeding the initial uplift of the Star Peak Range. He attributed the considerable vertical interval between the bottom of the zone of enrichment and the present position of the water table to the second and major uplift of the Star Peak Range. These conclusions are probably also applicable to the present area.

OXIDATION

Since the formation of the original supergene sulphide zone, erosion and lowering of the water table consequent upon uplift of the range have caused the zone of oxidation to encroach upon the sulphide zone. Oxidation products derived both from unreplaced hypogene sulphides and from supergene sulphides are therefore present in the sulphide ores. The enrichment by supergene native silver shown by some of the ores has presumably resulted from oxidation of silver-bearing sulphides, solution of the silver, and its reprecipitation at lower levels in the zone of oxidation.

ANTIMONY DEPOSITS

Mode of occurrence and form.—The principal antimony deposits of the area have been exposed in the Bloody Canyon and Black Warrior mines. A smaller deposit has been worked at the Pflum mine. The deposits at Bloody Canyon are part of a mineralized zone of steeply dipping sheet-fissure veins trending N. 30° W. to N. 25° E. The principal ore bodies appear to have been found on the south side of the south fork of the canyon in two veins, about 200 feet apart, in the uppermost flow of the rhyolite formation. The workings on the veins were only partly accessible at the time of visit, but the characteristics of the veins are well shown in a stope on the west vein.

In this stope the ore visible is largely confined to a zone 1 to 3 feet wide, but possibly thicker in mined-out portions. The zone strikes N. 5° E. to N. 5° W., dips 80° W., and is bounded by smooth, gouge-coated sheet surfaces. For the most part the "zone" is a simple quartz fissure vein, but in places inclusions of rhyolite are abundant, and the zone is a stockwork. Coarse, bladed stibnite, pure or intergrown with quartz, is distributed in streaks, bunches, and lenticular masses. The largest mass seen was 9 inches wide and 3 feet long, but the ore shoots stoped probably were larger lenses. For the most part stibnite has filled fractures in quartz or replaced fractured quartz, but at several places has formed veinlets in inclusions or even in fractures in the walls. Remnants of the vein visible in the workings indicate that the deposit consisted of productive shoots of high-grade ore alternating with nearly barren portions.

In both the productive veins, the ore bodies, so far as could be ascer-

tained, are confined to the uppermost rhyolite. In surface openings on the south end of the outcrop of the western vein the stopes end a few feet above the top of the rhyolite. The vein could not be traced across

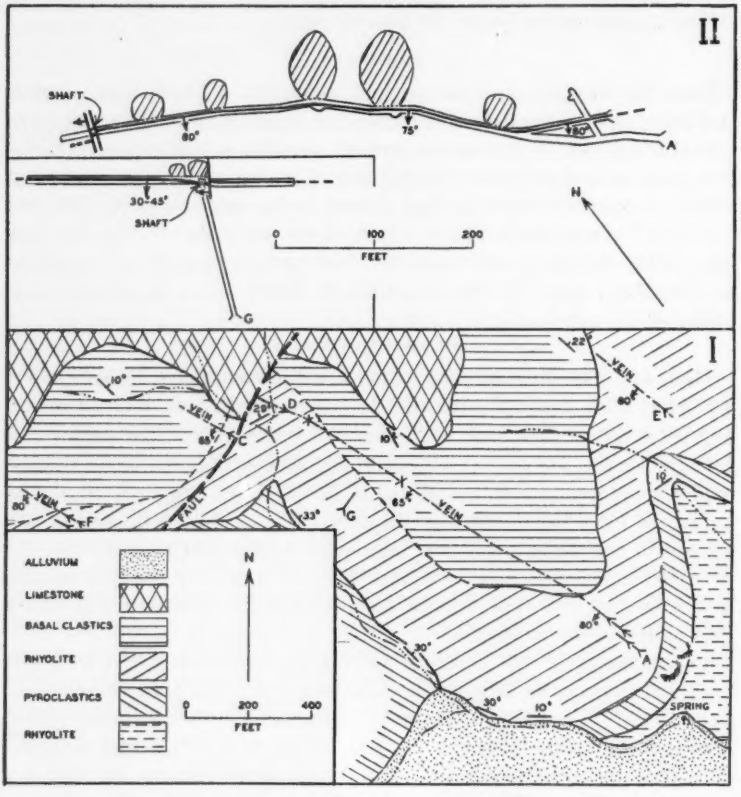

FIGURE 18.—*Geologic map and plans, Black Warrior mine*

I—geologic map of vicinity, based on pace-compass traverse. II—plans of workings on main vein at A and G of I.

the contact, but evidently the ore did not extend into the overlying basal clastics.

In the vicinity of the Black Warrior mine (Fig. 18) the strata dip eastward in general but are flexed into minor, northwest-trending folds. Roughly parallel to the folds are three fissure veins along narrow, steeply dipping sheeted zones. Most of the antimony ore was found in the

middle vein, traceable in scattered outcrops for about 1300 feet. In the principal tunnel (Fig. 18-II) the vein is in rhyolite. In places the vein is solid quartz and sulphides, a few inches to 3 feet in thickness; in places the vein splits into several branches along closely spaced, adjacent sheet planes. Smaller quartz veins parallel the main vein on either side.

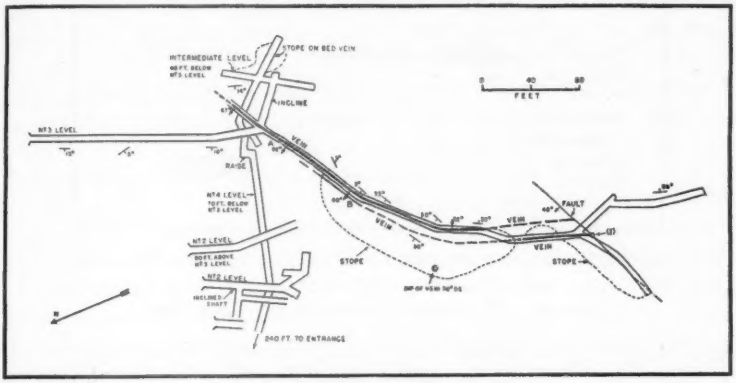

FIGURE 19.—*Plan of Pflum mine*
Based on survey by L. V. Pflum.

The vein contacts are frozen except where post-mineral movement has caused sheeting and brecciation of the vein parallel to the walls and development of gouge seams at the contacts. Faults which dip in general gently west or northwest and are marked by gouge seams offset the vein a few inches to 4 feet. A few are shown in Figure 18. The apparent displacement in each case is a strike-slip of the hanging wall to the south or southwest. A vein in the adit at C is probably the productive vein offset by one or more of these faults.

Unoxidized ore remaining at the tunnel level consists principally of milky, coarse, comby to massive quartz, in which stibnite, accompanied by traces of adularia, pyrite, sphalerite, and tetrahedrite, forms irregularly distributed streaks, bunches, and lenticular masses. The vein shows an alternation of ore-bearing and barren stretches, and the ore apparently formed flat lenticular shoots.

The ore bodies appear to be confined to the upper rhyolite. At two points shafts have been sunk on the productive vein, but examination of these was not feasible. None of the stopes seen extend below the tunnel levels. The veins at C, D, E, and F appear to have been unproductive of antimony.

In the Pflum mine granular stibnite is found in quartz fissure veins along faults of small displacement in the lower part of the Star Peak formation. The principal vein is 1 to 12 inches thick and occupies a rotational fault fissure of variable attitude (Fig. 19). The northwest wall is downthrown 1 foot at A, upthrown 3 feet at B. Part, at least, of this

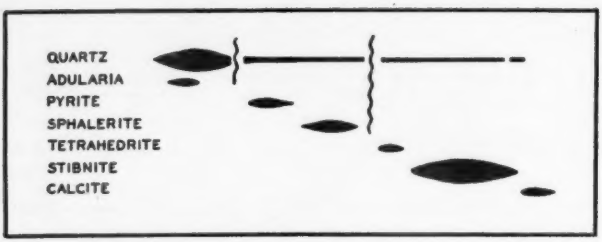

FIGURE 20.—*Sequence of hypogene mineral development in antimony ores*

Periods of deformation are indicated by vertical wavy lines.

movement was post-mineral, for at several places the vein is pinched out. A second vein, parallel to, and a few feet west of, the main vein is traceable to a point (Fig. 19-I) at which it appears to be cut off by the fault shown. The ore is milky, comby, or massive quartz, in which granular stibnite, intergrown with quartz and traces of sphalerite and pyrite, forms streaks, bands, and pockets. The ore minerals have replaced early-deposited vein quartz along fractures or have filled spaces produced by fracturing. There has been little replacement of silty limestone walls and inclusions.

Hypogene mineralization.—The hypogene mineral sequence shown by the veins is given in Figure 20. As in the silver ores, successive stages in mineralization can be recognized. In general, late minerals have replaced earlier minerals along fractures. The predominant mineral is coarse-grained milky quartz, massive or with comb structure, containing scattered quartz-lined vugs. Finely granular to coarsely bladed stibnite, generally intergrown with quartz, is irregularly distributed in streaks or bunches or in large lenticular shoots which furnished most of the commercial ore. Other hypogene minerals are present only in small amounts. Tetrahedrite (freibergite?) is present only in ore from the Black Warrior mine. At the Pflum mine freibergite, together with silver-bearing jamesonite, is present in bed-vein ore, but neither mineral has been found in the fissure-vein stibnite ore. On the other hand, no stibnite has been found in the bed-vein ore. Adularia, though present in altered wall

rocks at all the deposits, has been found in vein material only in ore from the Black Warrior mine. Calcite is also present only in ore from the Black Warrior mine.

Oxidation.—Effects of oxidation are conspicuous at all the deposits, especially in portions of the Bloody Canyon and Black Warrior mines at, or close to, the surface. At these mines both ores and wall rocks are stained by limonite and by yellow antimony oxides. At the Black Warrior mine, according to Whitehill (1875, p. 49), the ore from the upper part of the vein consisted of antimony oxides. At tunnel level the ore is in part unoxidized, but particularly in brecciated parts of the vein snow-white stibiconite is abundant. Associated with stibiconite is yellow chalcedonic quartz, apparently supergene, which in thin section shows relict "colloform" and pisolitic structures presumably indicating colloidal deposition. Valentinite, present in small amounts, shows replacement by stibiconite. The ore also contains traces of supergene covellite and is stained by limonite. Valentinite and stibiconite have also been recognized in thin sections of ore from the Bloody Canyon mine.

Quartz-stibnite veins similar to those of the present area have been reported from the Battle Mountain district, Nevada (Hill, 1915, p. 76), from the Yellow Pine district, Nevada (Ross, 1926), from Alaska (carrying gold) (Brooks, 1916, p. 10-11), and from other localities. Nearer at hand the Sutherland mine (Lincoln, 1923, p. 201-202) near the north end of the Humboldt Lake Range was an important source of antimony during the World War.

GOLD DEPOSITS

A number of the ores of the area contain gold in small quantities, but few deposits have been worked primarily for their gold content, and none, so far as known, has yielded a large production. The Marigold group of veins in Wilson and Straight canyons and the Tehama deposit in Tehama Canyon were examined. The hypogene mineralization processes at these two deposits are summarized in Figure 21. As in other ores of the area successive stages of mineralization can be recognized in the gold ores by the relations of various vein minerals to successive periods of fracturing. Scaly aggregates of biotite formed by replacement of orthoclase in rhyolite at the Tehama Mine are of interest as indicating that during early stages of mineralization temperatures were somewhat higher at the deposit than at most of the ore deposits of the area. The gold of small placer deposits found at various places in the canyons of the area is derived partly from gold-bearing veins such as those here described but in part from gold present in small amounts in the silver ores of the area.

The Marigold group of veins is a system of quartz fissure veins which

strike N. 20°–35° W., dip 70° W. to nearly vertical, and are enclosed in rhyolitic volcanics and granite porphyry on the south side of Straight Canyon (Pl. 1). The principal vein of the system, exposed in the workings of the Marigold mine, strikes N. 20° W. and dips 70° W. The vein

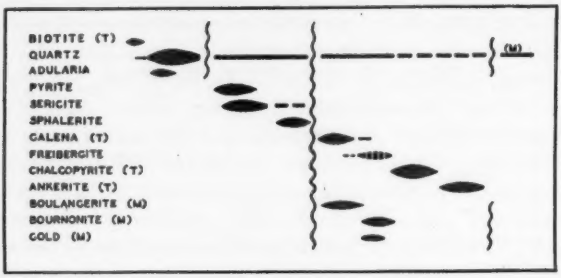

FIGURE 21.—*Sequence of hypogene mineral development in gold ores*

Periods of fracturing are indicated by vertical wavy lines. (T) = mineral occurring only in ore from the Tehama mine; (M) = mineral occurring only in ores from the Marigold group of deposits.

has a maximum width of 1 foot where seen. The wall rock is limonite-stained, sericitized, silicified, pyritized rhyolite. Quartz veinlets in the rhyolite contain considerable adularia. Both vein and adjacent wall rock are sheeted and shattered, and at the contacts there are thick seams of limonite-stained, clay-like quartz-sericite gouge, containing fragments of vein quartz and rhyolite. The gouges are bounded by gently curving sheet surfaces. Partial replacement of wall-rock inclusions is shown in some parts of the vein.

Another vein, 1 foot to 3 feet thick, exposed in workings at the forks of Straight Canyon, is similar to the Marigold vein but is enclosed in granite porphyry. This vein strikes N. 35° W. and dips 80° W.

Ore from these two veins differs significantly from that of other fissure veins only in metallic content. In unoxidized ore gold is apparently later than boulangerite, but its relation to bournonite is unknown. A little gold is present in partly oxidized ore from the upper portion of the vein in Straight Canyon. This ore shows hypogene sulphides partly replaced by supergene argentite, covellite, and chalcocite, and complex mixtures of cerussite, malachite, azurite, and antimony and manganese oxides. The origin of the gold is uncertain.

The Tehama deposit in Tehama Canyon (Pl. 1, A) is a sulphide-bearing quartz stockwork formed by mineralization of a zone of severe sheeting and brecciation in rhyolite. The rhyolite has been silicified, sericitized,

and pyritized. The principal mineralized fractures strike N. 20° E. and dip 70° W. According to local report the deposit was worked for gold, but the mineral is not present in specimens examined by the writer.

The ore is shattered rhyolite minutely veined and partly replaced by

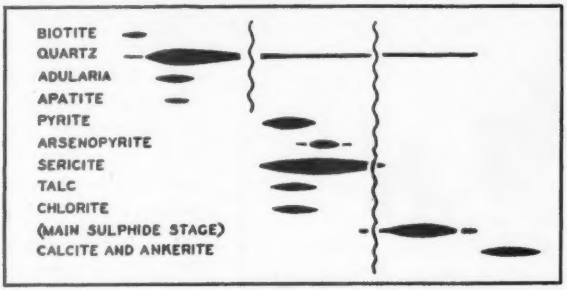

FIGURE 22.—*Sequence of development of hypogene products of wall-rock alteration*

Periods of fracturing indicated by vertical wavy lines.

sulphide-bearing quartz containing adularia. The principal sulphides appear to be chalcopyrite and pyrite, but sphalerite, freibergite, and galena are abundant. Ankerite is present as abundant disseminated rhombs. The portion of the deposit accessible is all within 100 feet of the surface. The ore shows enrichment by supergene chalcocite, with some covellite, and partial oxidation has resulted in abundant chalcanthite and limonite, with a little copiapite. Chalcanthite and an unidentified copper sulphate are abundant as efflorescences. Oxidation is particularly prominent near minor post-mineral slips of variable attitude, where the ore is shattered.

LEAD DEPOSIT

A quartz vein striking N. 30° E. and dipping 35° W., along a sheeted zone in pyritized, silicified, and sericitized rhyolite, is exposed to a depth of 75 feet in a shaft on the north side of Unionville Canyon (Pl. 1, I). The deposit is interesting chiefly because the only abundant sulphides are pyrite and galena. The ore is brecciated, and the sulphides in large part have been oxidized, chiefly to cerussite and limonite. The enclosing brecciated rhyolite contains cerussite in veinlets. According to local report the deposit was prospected for lead.

ORE FORMATION

WALL-ROCK ALTERATION

General statement.—Stages and products of wall-rock alteration inferred from thin-section studies are summarized in Figure 22.

Study of the wall rocks at the deposits, supplemented by petrographic studies, has shown that a considerable variety of alteration products has been formed during hypogene mineralization. However, no essential relationship between types of deposits—silver, gold, or antimony—and types of alteration has been disclosed. Differences in degree and kinds of alteration produced correspond rather to differences in the original compositions of various wall rocks. On this basis, two types of alteration are readily distinguishable: (1) Alteration of feldspathic rocks; (2) alteration of calcareous rocks.

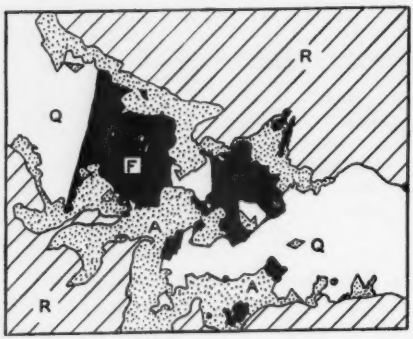

FIGURE 23.—*Freibergite* (*F*) *replacing quartz* (*Q*) *and adularia* (*A*)

Veinlet in silicified rhyolite (R). Ore from Tehama mine. Line drawing from photomicrograph. × 40.

Alteration of feldspathic rocks.—The term "feldspathic rocks" is here used to designate rhyolitic volcanics, granite porphyry, and tuffaceous basal clastics of the Star Peak formation. These rocks have been so altered by the mineralizing solutions that locally the original features of the rocks are obscured or destroyed, and boundaries of rock units can only be approximated. Sericitization and silicification were the chief alteration processes. At most of the deposits the earliest stage was silicification attended by formation of quartz veinlets containing small amounts of adularia (Fig. 23). Silicification apparently was long continued and was the predominant alteration process close to the veins. Wall rocks adjacent to or composing inclusions in the veins are generally heavily impregnated with replacing quartz and traversed by quartz veinlets. Replacement quartz is usually finer in texture than quartz forming fracture fillings, but both vary.

In general sericitization was subordinate adjacent to the veins but predominant at greater distances. Sericite composes disseminated flakes and fine aggregates or coarser streaks marking shear surfaces. Chlorite has formed only locally. Talc is abundant in tuffaceous strata at the Sheba and American Basin mines. Fine biotite aggregates have replaced orthoclase in rhyolite at the Sheba and Tehama deposits. Pyrite, with arsenopyrite at the Sheba and Champion deposits, is widely disseminated. Calcite is the most abundant carbonate. It is invariably a late mineral and in part may be supergene. At the Tehama deposit disseminated ankerite

replaces all other hypogene minerals except sulphides, both in wall rock and vein material.

Alteration of calcareous rocks.—In general the calcareous strata (limestones and silty limestones) are less altered than the feldspathic rocks. Veinlets of quartz, some of which contain traces of adularia and apatite, are numerous, but replacement of these rocks by quartz is largely confined to the borders of the channels of mineralization. Contacts between vein quartz and calcareous strata are generally sharp, even when marginal relations indicate replacement. Silicification of calcareous wall rocks has been observed only at a few places. It is a minor feature, except perhaps at the Sheba-De Soto deposit, where bands of cherty silica in the lower limestone strata may be due to hydrothermal silicification. Quartz formed by replacement of calcareous rocks is mostly of fine texture.

Disseminated pyrite is locally abundant adjacent to the veins. In the Sheba mine talc and dolomite form veinlets in certain limestone beds.

STRUCTURAL CONTROL OF MINERALIZATION

Distribution of deposits.—A close relationship to structural features characterizes the deposits. This relationship is manifested in the distribution of the deposits, in their forms, and in the distribution of the metallic minerals. Most of the deposits are found adjacent to or within major fault zones (Pl. 1). One group is largely confined to the down-faulted axial portion of the great anticline of the southern part of the area. Another consists of deposits along the fault zone which extends northward from American Basin. The Champion deposit is part of a mineralized zone within the great fault zone traversing upper Star Basin. The stibnite deposits in Bloody Canyon also adjoin a fault zone.

In view of the modes of occurrence of deposits within these zones, it seems improbable that this association is fortuitous. In the Pflum and Sheba mines veins fill pre-mineral fault fissures of small displacement. The Santa Clara, Tehama, and Champion deposits are parts of mineralized zones of fracturing associated with and trending roughly parallel to fault zones. A few veins, like the Governor Bradley vein (Pl. 1), occupy fault fissures of considerable displacement. These features and the evident dependence of hypogene mineralization throughout the area on channels produced by fracturing would seem to indicate that the major fault zones were in existence at the time of mineralization and that deformation attending faulting produced the channels along which the mineralizing solutions rose.

Forms.—The forms of the deposits closely reflect variations in fracture patterns from one rock unit to another and even within the same rock

unit. Consequently, although in some deposits a single form is persistent, in others gradual or abrupt transitions from one type to another are found. Where a deposit can be traced across boundaries between structurally dissimilar rock units, striking variations in form are found (Figs. 10 and 11). It follows that a simple grouping of the deposits on the basis of form is not feasible, although for convenience of description bed veins, fissure veins, and stockworks have been distinguished.

In massive rocks, particularly in rhyolite and granite porphyry, the typical deposits are well-defined fissure veins along steeply dipping sheet fractures. Narrow stockworks are found as local phases of the fissure veins where the sheet fractures are branched and cross-fracturing of the included sheets has taken place. Larger stockworks, like the Tehama and Champion deposits, occupy broader zones of sheet fissures connected by cross-fractures.

The extensive quartz bed veins in the lower part of the Star Peak formation are striking illustrations of structural control of form. The bed-vein form is primarily ascribable to separation of certain silty limestone strata along closely spaced lamination planes. The extraordinary persistence with which veins of this type have followed particular horizons in the formation is emphasized by the fact that parallel to the main vein in the south workings of the Arizona mine there is a 1-inch bed vein which maintains its thickness, continuity, and position below the main vein for a known distance of 300 feet.

It is striking that in spite of intensive prospecting in the area no workable deposit in the Star Peak formation more than a few hundred feet above its base is known, although the outcrop area of the upper members is large. The cause would seem to have been lateral spreading of the mineralizing solutions along structurally favorable horizons in the lower part of the Star Peak formation. This spreading is clearly shown in the existence of bed veins but is indicated even where fissure veins and stockworks have formed in the lower part of the formation. In the Sheba deposit the great importance attached by the early miners to the shaly limestone strata known as the "hanging wall" was a recognition of the fact that these strata apparently constituted a relatively impervious barrier beneath which the mineralizing solutions spread.

Modification of forms by replacement.—Although it is evident that structural features have largely determined the forms of the deposits, replacement of wall rocks took place during mineralization. This is shown by familiar phenomena of "ghosts" of replaced wall-rock inclusions and at many places by marginal relations, megascopic or microscopic, between vein matter and wall rocks. Replacement was erratic, and differences in

extent of replacement are shown even within a single deposit. In general, however, replacement was subordinate to fracture filling, and its effect was merely to modify forms essentially determined by fracturing.

Distribution of metallic minerals.—The hypogene metallic minerals were deposited chiefly by replacement of early-formed quartz along channels produced by fracturing. Where the veins were sheeted parallel to the walls streaky or banded ore has resulted. Where the fracture pattern was irregular, the metallic minerals are irregularly distributed in pockets and bunches.

Post-mineral deformation.—All the deposits show effects of post-mineral deformation. Contacts of veins with wall rocks are frequently marked by seams of gouge, and the whole width of a vein may consist of gouge containing angular fragments of quartz. Post-mineral faults are numerous at some of the deposits.

GENESIS OF HYPOGENE ORES

Genetic relationships among deposits.—The hypogene deposits of the area were formed along fractures of various patterns in rhyolitic volcanics, granite porphyry, and the lower strata of the Star Peak formation. Mineralization was effected by fracture filling and replacement and entailed rather extensive alteration of the enclosing wall rocks.

The basis on which the ore deposits are grouped in the present report is essentially economic—*i. e.*, according to whether gold, silver, or antimony was present in minable quantities. Such a classification does not correspond to any sharply defined natural divisions and has been adopted because natural divisions are not suggested by the deposits. Though differing considerably in detail, the deposits have significant features in common. A comparison of Figures 12, 20, 21, and 22 indicates that at all the deposits mineralization took place by successive stages separated by fracturing. During the first stage coarse milky quartz was introduced, together with small amounts of adularia at most of the deposits. During the second stage quartz and adularia were fractured and partly replaced, successively, by pyrite, arsenopyrite (at certain deposits), and sphalerite. Renewed fracturing opened a third stage, during which the remainder of the hypogene metallic minerals, including silver and antimony sulphides and gold, were deposited. In certain ores deformation during the second and third stages is indicated. Quartz apparently was deposited during all stages. A significant point is that the hypogene minerals of the valuable metals were all deposited at late stages in mineralization, subsequent to emplacement of the bulk of the vein material.

On a mineralogical basis the stibnite deposits are the most clearly

defined group. Yet stibnite is present in a few specimens of ore from the Sheba mine, and sphalerite and pyrite, together with tetrahedrite at the Black Warrior mine, in stibnite ore. The hypogene mineralogy of the veins has no apparent relation to any particular form of deposit or type of wall rock.

The ore deposits are further similar in the character of the alteration of the associated wall rocks. Silicification, pyritization, and sericitization were everywhere the principal alteration processes. Except at the Sheba mine adularia is present either in the ores or in veinlets in the wall rocks. Differences in mode of alteration appear to be related to local variations in temperature and in compositions of wall rocks, rather than to significant differences in the characters of the mineralizing solutions.

Mineralogically, texturally, and structurally the hypogene deposits are a closely related assemblage. No grounds, structural, stratigraphic, or mineralogical, for distinguishing two or more epochs of mineralization separated by significant time intervals have been discovered. It is therefore concluded that the deposits were formed during a single epoch of mineralization and probably within a brief interval of geologic time. Only with respect to the later stages of mineralization can significant differences be recognized.

Time and source of mineralization.—With regard to the time of mineralization, all that is known from the present area is that mineralization followed consolidation of post-Jurassic (?) intrusives, preceded the intrusion of Tertiary olivine diabase dikes, and was guided by structural features which appear related to Jura-Cretaceous (?) folding. The widespread and presumably contemporaneous mineralization in the southern part of the range, however, has been found by Knopf and Jenney and Kerr to be genetically related to the Jura-Cretaceous (?) granitic intrusives. Knopf (1924, p. 35) suggested that the source of the mineralizing solutions was the parent magma of the aplite, although he recognized that mineralization has affected later granite porphyry. Jenney (1935, p. 42), however, referred a large amount of the mineralization to the Rocky Canyon granite, which has invaded granite porphyry and is itself hydrothermally altered. Kerr (1938, p. 403) found the tungsten mineralization at Oreana to be closely related to late aplite dikes.

Conditions during mineralization.—In the Rochester district Knopf (1924, p. 57) distinguished three principal types of deposits: (1) Dumortieritic deposits, carrying only sporadic tourmaline and no sulphides; (2) tourmaline-gold-quartz deposits with coarse microcline considered to indicate temperatures of 400°–600° C.; (3) silver veins, in part tourmalinic. Transitions between types were not found. From the western side of the

range north of the Rochester district, Jenney described a pneumatolytic stage beginning with infiltration of quartz-albite end-stage material derived from the Rocky Canyon granite mass and further marked by development of andalusite, biotite, and tourmaline, with other minerals. The deposits described by Knopf were referred by Jenney (1935, p. 43) to a succeeding hydrothermal stage.

By contrast with the areas to the south the present area is characterized by the slight extent to which high-temperature phases of mineralization are recorded. Tourmaline has been observed only in the altered diabase of the range crest. Quartz-albite material forming veinlets in tuffaceous strata near the Arizona mine and replacing potash feldspars of the granite porphyry masses may represent a high-temperature phase of mineralization. However, the earliest stage of mineralization in the present area was marked at most of the ore deposits by development of adularia, which suggests temperatures between 100° and 300° C. (Grout, 1932, p. 441; Lindgren, 1933, p. 454; Koenigsberger, 1901, p. 107). Only at the Tehama and Sheba deposits does replacement of orthoclase by biotite suggest somewhat higher temperatures. During sulphide deposition the temperatures must have been lower, as indicated by the finer texture of quartz deposited at this stage. The absence of high-temperature deposits in the present area corresponds to a smaller development of post-Jurassic (?) granitic intrusives and would therefore seem to confirm the hypothesis that the source of the mineralization was the parent magma which gave rise to the granitic intrusives extensively exposed to the south.

In the Rochester district Knopf (1924, p. 57) noted the occurrence within small areas of all three principal types of deposits recognized. He says:

> "The conclusion adopted is that the three types of deposits owe their origin to solutions of distinctive character, which were given off successively by the consolidating magma."

A similar phenomenon is exhibited in the present area, although the contrasts between types are less marked. However, the similarity of early mineralization stages shown by deposits of the present area suggests that early quartz and adularia were formed at about the same time in all the deposits. It is here suggested that on the basis of Knopf's hypothesis differences in subsequent mineralization in the present area might be explained by differences in the times at which renewed deformation produced channels of sulphide mineralization at different deposits by fracturing of early-deposited quartz. Progressive changes in the composition of solutions given off by the source magma would then result in variations in the metallic content of the ores deposited.

Most of the deposits in the present area are near the contact between the rhyolite and the Star Peak formation. From available information

the thickness of overlying strata at the time of mineralization would appear to have been between 5000 and 7000 feet, on the presumption that mineralization took place before the post-Jurassic erosion interval was well advanced.

The silver deposits of the Humboldt Range have been assigned by McKnight (1933, p. 585) to the quartz-tetrahedrite-galena group of mesothermal deposits as defined by Lindgren. This classification would appear to be justified in general by the structural and mineralogical characteristics of the silver deposits of the present area. Adularia in mesothermal deposits is unusual (Lindgren, 1933, p. 530) but not unknown, for at the Premier mine (Burton, p. 588-589, 596) British Columbia, considered mesothermal by Lindgren (1933, p. 566), adularia is an abundant gangue mineral.

Certain features of the deposits of the area, however, suggest that the deposits should be assigned to the "leptothermal" zone, proposed by Graton (1933) to distinguish those deposits which are neither of epithermal nor of typically mesothermal types. The hypogene ore minerals include both simple sulphides and more complex sulpho-salts of the base metals and of silver. Gold is subordinate. Quartz is the dominant gangue mineral and is accompanied by adularia. The structural and textural features of the deposits and the indicated depth of formation also agree fairly well with the characteristics of the leptothermal group as described by Graton.

Schrader (1933, p. 659) has classified the antimony deposits of the present area as epithermal. The results of the present study, however, indicate that the antimony deposits and the silver deposits were formed under essentially similar conditions, although stibnite, which is later than associated hypogene sulphides where observed, may well have been deposited at lower temperatures than those which prevailed during the main silver mineralization.

SUMMARY OF CONCLUSIONS

The principal conclusions which have resulted from the present investigation of the northeastern part of the Humboldt Range may be summarized as follows:

(1) The oldest rocks in the area are a series of late Paleozoic or lower Triassic keratophyric volcanics. The characteristic members of this series are rocks which have been produced by shearing and hydrothermal alteration of extrusives which probably ranged originally from trachyte to olivine basalt.

(2) Albite in the keratophyres and in the post-Jurassic basic intrusives has been produced from anorthite-bearing plagioclase by shearing

and hydrothermal alteration attending Jura-Cretaceous (?) deformation and mineralization.

(3) Overlying the keratophyric volcanics is a series of altered rhyolitic volcanics which are equivalent to part or all of the Rochester-Weaver sequence of volcanics described by Knopf and Jenney from the southern portion of the Star Peak Range.

(4) The section of the Star Peak limestone as given by King and Smith is incorrect, owing to structural complications not recognized by these investigators. The formation is here redefined as comprising Middle and Upper Triassic sediments overlying the rhyolite formation. The total thickness of the formation is provisionally estimated at 4500 to 6500 feet of which the lower 3500 feet is exposed in the present area.

(5) The Jura-Cretaceous(?) intrusive sequence of the Star Peak Range includes, in addition to types recognized by previous investigators, olivine diabase, gabbro porphyry, and syenite.

(6) The principal ore deposits of the northeastern Humboldt Range are deposits of silver, gold, and antimony, found in fissure veins, stockworks, and bed veins, in rhyolite, in granite porphyry, and in the lower part of the Star Peak formation.

(7) The areal distribution and the forms of the deposits and the distribution of the minerals within them show a close control by deformation prior to and during mineralization.

(a) A close relationship exists between the distribution of the deposits and the occurrence of, and structural features of, major fault zones in the area. This relationship is believed to indicate that the channels of mineralization were produced by the same deformation which produced the fault zones, and that the fault zones, therefore, were initiated prior to mineralization.

(b) Variations in the forms of the deposits are due principally to differences in structural features produced by deformation in different types of wall rocks. Replacement of wall rocks has taken place, but its effect has been merely to modify forms which have been essentially determined by fracturing.

(c) Development of the ores proceeded in fairly well defined stages delimited by successive periods of fracturing during mineralization. Deposition proceeded by successive metasomatic replacements and was guided by fractures produced by the successive deformations.

(8) The close relationship between mineralization and deformation shown in the deposits of the area suggests that the development of different types of deposits may be due to differences in the times at which renewed deformation produced channels of sulphide deposition in early-deposited vein quartz.

(9) Mineralogically, texturally, and structurally the hypogene deposits of the area are closely related and have evidently formed during a single epoch of mineralization under essentially similar conditions of moderate temperature and pressure.

10. The absence of high-temperature deposits in the area, together with the fact that post-Jurassic granitic intrusives are much less widely developed than in the southern portion of the range, appears to confirm the hypothesis, advanced by earlier investigators, that mineralization was effected by solutions genetically related to the granitic magmas.

(11) The development of commercial silver ores, at least in most of the deposits, is due to enrichment of lean sulphide ores by supergene argentite, native silver, and possibly cerargyrite. Formation of supergene argentite, and associated covellite and chalcocite, preceded development of the present topography.

WORKS TO WHICH REFERENCE IS MADE

Bailey, E. B., and Grabham, G. W. (1909) *Albitization of basic plagioclase feldspars,* Geol. Mag., vol. 6, p. 250-256.

Blackwelder, E. (1931) *Pleistocene glaciation in the Sierra Nevada and Basin ranges,* Geol. Soc. Am., Bull., vol. 42, p. 865-922.

Brooks, A. H. (1916) *Antimony deposits of Alaska,* U. S. Geol. Survey, Bull. 649.

Burton, B. S. (1868) *Contributions to mineralogy,* Am. Jour. Sci., 2d ser., vol. 45, p. 36-38.

Burton, W. D. (1926) *Ore deposition at the Premier mine, British Columbia,* Econ. Geol., vol. 21, p. 586-604.

Colony, R. J. (1923) *The final consolidation phenomena in the crystallization of igneous rock,* Jour. Geol., vol. 31, p. 169-178.

Daly, R. A. (1933) *Igneous rocks and the depths of the earth,* p. 421.

Dewey, Henry, and Flett, J. S. (1911) *Some British pillow-lavas and the rocks associated with them,* Geol. Mag., vol. 8, p. 202-209.

Emmons, W. H. (1933) *Recent progress in studies of supergene enrichment,* in *Ore deposits of the Western States,* p. 386-418.

Gabb, W. M. (1864) *Description of the Triassic fossils of California and adjacent territories,* Calif. Geol. Survey, *Paleontology,* vol. 1, p. 17-35.

Gilbert, G. K. (1928) *Studies of Basin Range structure,* U. S. Geol. Survey, Prof. Paper 153.

Gilluly, James (1935) *Keratophyres of eastern Oregon and the spilite problem,* Am. Jour. Sci., vol. 29, p. 225-252, 336-352.

Graton, L. C. (1933) *The hydrothermal depth-zones,* in *Ore deposits of the Western States,* p. 181-197.

Grout, F. F. (1931) *Petrography and Petrology,* p. 441.

Hague, Arnold, and Emmons, S. F. (1877) *Descriptive geology,* Rept. Geol. Expl. 40th Par., Prof. Papers Eng. Dept., U. S. Army, no. 18, vol. 2.

Hague, J. D. (1870) *Mining industry,* Rept. Geol. Expl. 40th Par., Prof. Papers Eng. Dept., U. S. Army, no. 18, vol. 3.

Hill, J. M. (1915) *Some mining districts in northeastern California and northwestern Nevada,* U. S. Geol. Survey, Bull. 594, p. 76-77.

Jenney, C. P. (1935) *Geology of the central Humboldt Range, Nevada,* Univ. Nevada, Bull., vol. 29, no. 6.

Johnston, W. D., Jr. (1938) *Vein-filling at Nevada City, California,* Geol. Soc. Am., Bull., vol. 49, p. 23-34.

Jones, J. C. (1913) *Geology of Rochester, Nevada,* Min. Sci. Press, vol. 106, p. 737-738.

Kerr, P. F. (1938) *Tungsten mineralization at Oreana, Nevada,* Econ. Geol., vol. 33, no. 4, p. 390-427.

———, and Jenney, C. P. (1935) *The dumortierite-andalusite mineralization at Oreana, Nev.,* Econ. Geol., vol. 30, p. 287-300.

King, C. E. (1878) *Systematic geology,* Rept. Geol. Expl. 40th Par., Prof. Papers Eng. Dept., U. S. Army, no. 18, vol. 1.

Knopf, A. (1924) *Geology and ore deposits of the Rochester district, Nevada,* U. S. Geol. Survey, Bull. 762.

Koenigsberger, Johannes (1901) *Die Minerallagerstaetten im Biotitprotogin des Aarmassivs,* Neues Jahrb., Beil. Band 14, p. 43-119.

Lindgren, Waldemar (1933) *Mineral deposits,* 4th ed., revised, 797 pages.

Lincoln, F. C. (1923) *Mining districts and mineral resources of Nevada,* 295 pages.

Longwell, C. R. (1930) *Faulted fans west of the Sheep Range, southern Nevada,* Am. Jour. Sci., ser. 5, vol. 20, p. 1-13.

Louderback, G. D. (1904) *Basin Range structure of the Humboldt region,* Geol. Soc. Am., Bull., vol. 15, p. 289-346.

Meek, F. B. (1877) *Paleontology,* Rept. Geol. Expl. 40th Par., Prof. Papers Eng. Dept., U. S. Army, no. 18, vol. 4, pt. 1.

Merriam, J. C. (1902) *Triassic Ichthyopterygia from California and Nevada,* Univ. Calif. Publ., Bull., Dept. Geol., vol. 3, no. 4, p. 63-108.

——— (1905) *A primitive ichthyosaurian limb from the Middle Triassic of Nevada,* Univ. Calif. Publ., Bull., Dept. Geol., vol. 4, no. 2, p. 33-38.

——— (1908) *Triassic Ichythyosauria, with special reference to the American forms,* Univ. Calif., Mem. 1, p. 1-196.

——— (1910) *The skull and dentition of a primitive ichthyosaurian from the Middle Triassic,* Univ. Calif. Publ., Bull., Dept. Geol., vol. 5, no. 24, p. 381-390.

McKnight, E. T. (1933) *Mesothermal silver-lead-zinc deposits,* in *Ore deposits of the Western States,* p. 582-602.

Ransome, F. L. (1909) *Notes on some mining districts in Humboldt County, Nevada,* U. S. Geol. Survey, Bull. 414. 75 pages.

Raymond, R. W. (1869) *Mineral Resources of the states and territories west of the Rocky Mountains,* 256 pages.

——— (1872) *Statistics of mines and mining,* 566 pages.

——— (1873) *Statistics of mines and mining* (for 1871), 566 pages.

——— (1875) *Statistics of mines and mining,* 540 pages.

——— (1877) *Statistics of mines and mining,* 519 pages.

Ross, C. P., and Schrader, F. C. (1926) *Antimony and quicksilver deposits in the Yellow Pine district, Idaho,* U. S. Geol. Survey, Bull. 780, p. 144-156.

Russell, I. C. (1885) *Geologic history of Lake Lahontan,* U. S. Geol. Survey, Mon. 11, 288 pages.

Schneiderhoehn, Hans, and Ramdohr, Paul (1931) *Lehrbuch der Erzmikroskopie,* Band 2, 714 pages.

Schrader, F. C. (1914) *The Rochester mining district, Nevada,* U. S. Geol. Survey, Bull. 580, p. 325-372.

Smith, J. P. (1904) *The comparative stratigraphy of the marine Trias of western North America,* Calif. Acad. Sci., Pr., ser. 3, vol. 1, p. 323-430.

——— (1914) *The Middle Triassic marine invertebrate faunas of North America,* U. S. Geol. Survey, Prof. Paper 83, 254 pages.

——— (1927) *The Upper Triassic marine invertebrate faunas of North America,* U. S. Geol. Survey, Prof. Paper 141, 262 pages.

———, and Hyatt, A. (1905) *The Triassic cephalopod genera of America,* U. S. Geol. Survey, Prof. Paper 40, 394 pages.

Sperb, Grace (1937) *An invertebrate fauna from the lower Star Peak formation of the West Humboldt Range, Nevada,* Unpublished master's thesis, Columbia University.

Stretch, R. H. (1867) Annual report of the State Mineralogist of Nevada for 1866.

Wheeler, H. E. (1936) *Helicoprion in the Anthracolithic of California and Nevada, and its stratigraphic significance,* Geol. Soc. Am., Pr. 1935, p. 394.

——— (1938) *Helicoprion in the Anthracolithic of California and Nevada, and its stratigraphic significance,* Manuscript of paper now in press. Used by permission of the author.

Whitehill, H. R. (1873) *Biennial report of the State Mineralogist of the State of Nevada, for the years 1871 and 1872.*

——— (1875) *Biennial report of the State Mineralogist of the State of Nevada, for the years 1873 and 1874.*

Zies, E. G., Allen, E. T., and Merwin, H. E. (1916) *Some reactions involved in secondary copper sulphide enrichment,* Econ. Geol., vol. 11, p. 407-503.

Zirkel, Ferdinand (1876) *Microscopical petrography,* Rept. Geol. Expl. 40th Par., Prof. Papers Eng. Dept., U. S. Army, no. 18, vol. 6.

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EXPLANATION OF PLATES 5 AND 6

PLATE 5

PHOTOMICROGRAPHS OF ROCKS

(1) Keratophyre, second spur south of Santa Clara Canyon. Albite phenocrysts in a dark groundmass with a faintly discernible pilotaxitic structure. × 25.

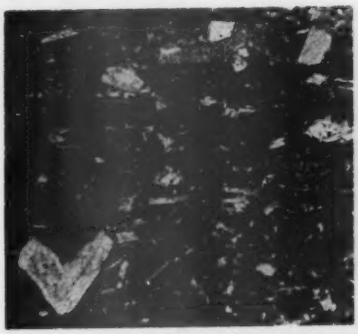
(2) Keratophyre tuff, second spur south of Star Canyon. Lithic fragment in a schistose, sericitic matrix. × 17.

(3) Rhyolite, south of Governor Bradley mine. "Chessboard" albite phenocryst in flow-banded groundmass, with streaks of introduced quartz parallel to flow structure. × 17.

(4) Sheared tuff, north of Bloody Canyon mine. Partly silicified felsite fragment in schistose, quartz-sericite matrix. × 16.

(5) Metagabbro porphyry, north rim of Star Basin. Zoned plagioclase laths fractured and partly replaced by secondary aggregates. × 19.

(6) Partly silicified granite porphyry, Arizona mine. Part of a large spherulite (dark area) traversed by quartz veinlets and partly replaced by quartz (light gray) adjacent to a quartz phenocryst enclosed in the spherulite. × 19.



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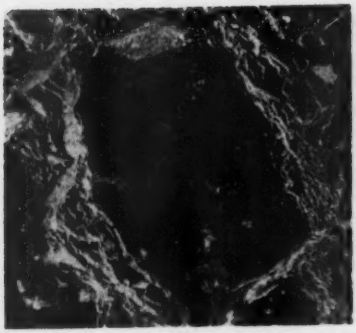

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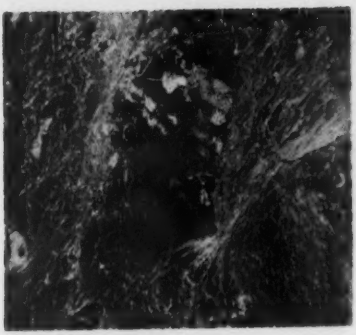

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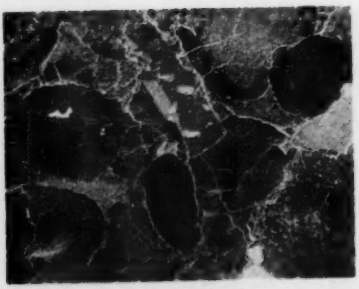
PHOTOMICROGRAPHS OF ROCKS

1



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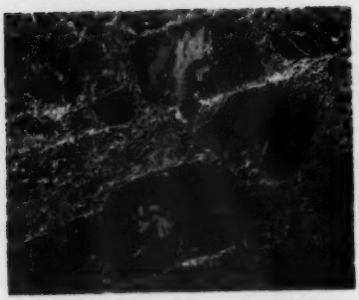

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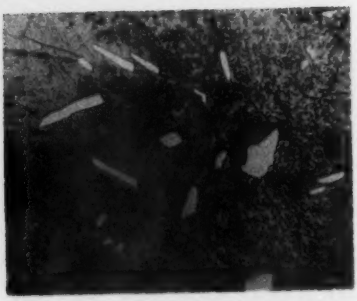

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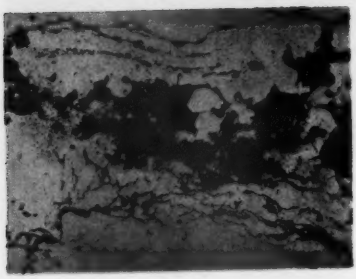

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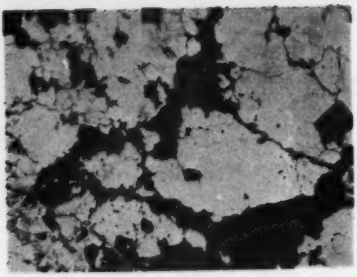


6

PHOTOMICROGRAPHS OF ROCKS AND ORES

PLATE 6

PHOTOMICROGRAPHS OF ROCKS AND ORES

(1) Altered granite porphyry, Arizona mine. Large phenocryst of orthoclase sericitized along fractures, phenocrysts of quartz (white), and sericitized spherulites in a fine quartz-sericite matrix. × 12.

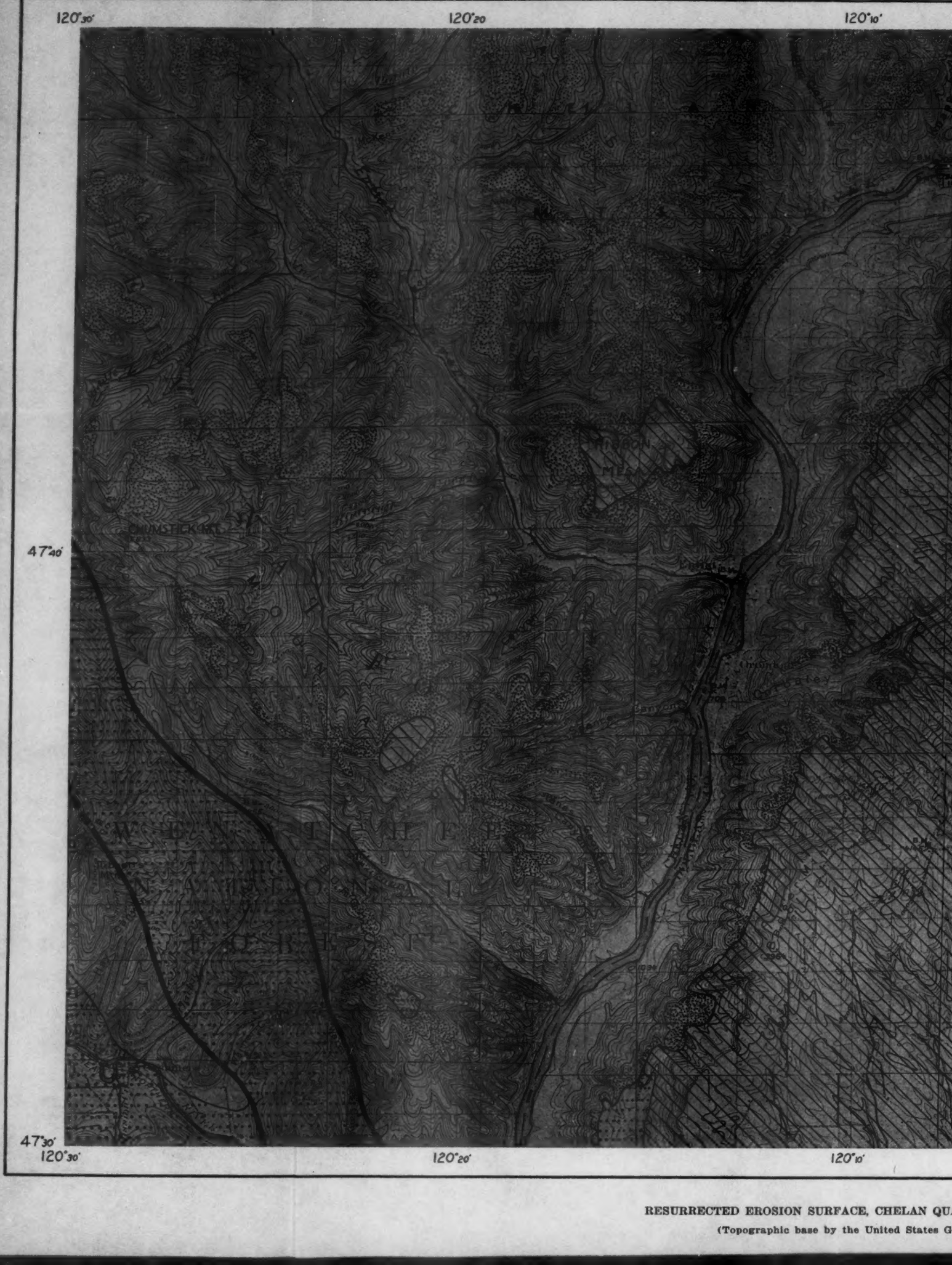
(2) Altered granite porphyry, Fourth of July Flat. Quartz (dark) and feldspar phenocrysts in a partly sericitized spherulitic groundmass. × 12.

(3) Altered rhyolite, Sheba mine. Orthoclase phenocrysts (dark) partly replaced by quartz (light gray), and rounded quartz phenocrysts (medium gray) in a schistose, sericitic matrix containing disseminated pyrite and arsenopyrite (black). × 14.

(4) Inclusions of silver-jamesonite(?) in galena (etched, gray). Sheba mine. × 106.

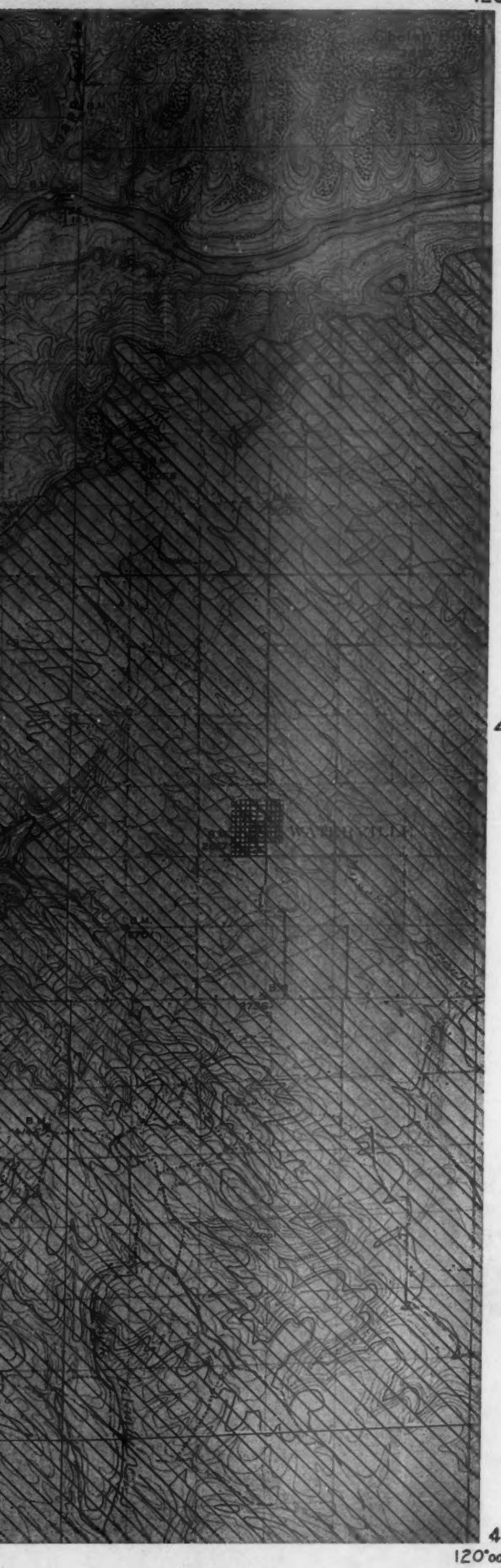
(5) Bed vein ore, Arizona mine. Bands and streaks of sulphides (black) formed by replacement of quartz (gray) along shear zones. × 13.

(6) Stockwork ore, Champion mine. Replacement of vein quartz by sulphides (black) along irregular fractures. × 13.



RESURRECTED EROSION SURFACE, CHELAN QU

(Topographic base by the United States G



LEGEND

RESURRECTED EROSION SURFACE

COLUMBIA RIVER BASALT
WITH ASSOCIATED SEDIMENTARY ROCKS AND PALAGONITE TUFFS

SWAUK FORMATION
CONGLOMERATE ARKOSE AND SHALE

GNEISS AND GRANITIC ROCKS
MAINLY THE SWAKANE GNEISS AND THE CHELAN GRANODIORITE BATHOLITH

NORMAL FAULT

SCALE
1 ½ 0 1 2 3 4 5 MI.

ELAN QUADRANGLE, WASHINGTON
d States Geological Survey)

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 50, PP. 635-660, 4 PLS. APRIL 1, 1939

RESURRECTED EROSION SURFACE IN CENTRAL WASHINGTON

BY A. C. WATERS

(With Discussion by Bailey Willis)

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ABSTRACT

Results obtained in mapping the geology of the Wenatchee-Chelan and of the Okanogan district suggest modifications of the earlier interpretations regarding the erosional history:

1) The existence of a "Methow Plain" or "Cascade Peneplain" is questioned. Most of the areas formerly considered to be remnants of this peneplain are surfaces controlled by the structure of the Columbia River basalt flows without regard to a widespread base level of erosion. One remnant, which bevels the Columbia River basalt and the Ellensburg formation on the flanks of an anticline in the Yakima district, is believed to be a pediment of very local extent.

2) The mature surface of the "Entiat stage" is well developed only on resistant granodiorite and gneiss and is exceedingly fragmentary or absent in areas underlain by greenstone, phyllite, schist, and Tertiary sedimentary rocks. It has not been recognized in areas underlain by the Columbia River basalt. The writer interprets nearly all the residuals of the "Entiat surface" described in the literature as resurrected fragments of a widespread late mature erosion surface buried beneath the Miocene Columbia River lavas and associated sediments and then partially exhumed in post-Miocene time. Certain fragmentary remnants of this surface on massive crystalline rocks beyond the limit reached by the basalt flows are believed to have been similarly preserved by burial under andesitic pyroclastics erupted from volcanoes in the Cascade Range.

INTRODUCTION

The groundwork for the geology of the Cascade Range was laid in the closing years of the last century by reconnaissance parties of the United States Geological Survey. At that time the concept of the peneplain, already firmly established by work on the Atlantic seaboard, was brought to the Cascade area by Bailey Willis and was adopted by his colleagues, I. C. Russell and G. O. Smith. The first statement in print of their application of this concept to the new area was made by Russell (1900, p. 144) who, impressed by the fact that the serrate alpine summits of the Northern Cascades integrate into a fairly level sky line when viewed from a high elevation, wrote:

> "The Cascade Mountains, as we know them, seem to have been carved from an upraised peneplain. This plain we term the Cascade peneplain, and the plateau may be conveniently designated the Cascade plateau."

The more detailed studies of Willis (1903) and of Smith (1903 a) were published a few years later. In the Wenatchee-Chelan district Willis was able to decipher a rather complex series of erosional events which he has conveniently summarized in the table reproduced below:

TABLE 1.—*Physiographic development of the Cascade Range* (Willis, 1903, p. 70)

Physiographic stage	Type locality	Nature of characteristic activity	Date, if one uninterrupted Glacial epoch	Date, if two or more Glacial epochs	Period
Stehekin..	Stehekin sources and valley.	Glacial retreat and reexcavation of old valleys.	Post-Glacial to present.	Post-Glacial to present.	Recent.
Chelan....	Gorge of Lake Chelan and terraces of the Columbia.	Glacial occupation of canyons, canyon modification, and valley filling by overloaded streams. Glacial diversion of streams.	Glacial epoch.	Latest Glacial epoch.	Pleistocene.
Twisp.....	Canyon of the Twisp, Methow quadrangle.	General acceleration of corrasion, resulting in excavation of canyons and ravines with readjustments of drainage.	Pre-Glacial..	Inter-Glacial.	
Entiat....	Basin of the Entiat and foot spurs of the Chelan Range.	Development of mature topography generally throughout the Cascade plateau.	Pre-Glacial..	Earlier Glacial epoch.	Pliocene or Pleistocene. (Sierran Age.)
Methow...	Generally throughout the broad mountain district, and specifically in Yakima Valley.	Planation by erosion to a low plain with monadnocks.	Pre-Glacial..	Pre-Glacial..	Pliocene.

In this table the "Methow stage" corresponds with the Cascade Peneplain, or Cascade Plateau, of Russell. Over most of the northern Cascades mere fragments of it were believed to have survived the later erosion, but in the Chelan quadrangle Willis (1903, Pls. 16, 17, 18) maps large residuals on the Waterville Plateau east of Columbia River and smaller ones in the mountains west of the river.

Much more conspicuous are the surfaces attributed to the "Entiat stage" of erosion. They are gently to steeply sloping upland surfaces, with an average relief of about 2100 feet, which are best seen in the long spurs at the eastern end of the Chelan Mountains and in the higher slopes adjacent to the lower course of the Entiat River. Willis attributes them to a stage of erosion which reduced the original Methow Plain to a rolling matureland.

According to Willis, renewed elevation occurred after the development of the Entiat matureland which resulted in the rapid incising of the streams to produce the canyons of the "Twisp stage." The pronounced topographic unconformity between the gently sloping surfaces of the Entiat stage and the sharp cliffs and abrupt slopes of the Twisp canyons is one of the most conspicuous topographic features of the district, giving rise to superlative examples of "two-story valleys." Willis considered the uplift that initiated the Twisp canyon-cutting to be complex and differential. Among its effects was the local accentuation of the northwest-southeast anticlinal ridges (Smith, 1903 b) which are such prominent features of central Washington. Local details of the Twisp surface are described by Willis (1903, p. 80-81), and his interpretation of it as the product of an episode of canyon-cutting immediately preceding glaciation is amply confirmed by later work.

Since the two youngest stages, the Chelan and Stehekin, relate only to the glacial features they need not concern us. Recent work [1] reveals that deposits left by at least three different glacial advances are present in the Wenatchee-Chelan area. Consequently the dates given in the second column of Willis' table apply to his interpretation of the age of the features of the district.

FIELD WORK AND ACKNOWLEDGMENTS

Investigation of the bedrock geology of the Wenatchee-Chelan district was begun by the writer in 1925 and has since been carried on at intervals. The completion of the work, made possible by a grant from the Geological Society of America, was accomplished during the summer of 1936.

[1] Detailed evidence of multiple glaciation in central Washington has been worked out mainly by Ben M. Page. His results are not yet published.

In the early stages of the mapping the writer accepted the interpretations of earlier workers with regard to the erosional history of the area. As the mapping progressed, however, new facts began to accumulate which were not consistent with the older conclusions. Most important of these new facts was the definite relationship of the residuals identified with the Methow and Entiat stages to restricted areas with a characteristic bedrock geology. Eventually, a new interpretation of the erosional development of the district consistent with both older and newer observations was conceived and is set forth in this report. Although the new interpretation differs radically from its predecessor it should not be deduced from this that the observations of the earlier workers were inaccurate or sketchy. On the contrary, central Washington has been particularly fortunate in the quality of its early reconnaissance geologic reports. The observations contained in them are reliable and amazingly abundant despite the short time at their authors' disposal and the primitive living and transportation conditions prevailing in this district near the end of the nineteenth century. The writer owes a personal debt to Willis and to Smith not only for their observations but also for the inspiration derived from the fact that work of such high calibre could be accomplished under the conditions in this district 40 years ago. Furthermore, it should be remembered that their work on the erosional history of the area was but one small phase of the broader geologic studies in which they were engaged.

The writer is indebted to Ben M. Page and to Konrad Krauskopf for critical reading of the manuscript.

BEDROCK GEOLOGY

The Wenatchee-Chelan district lies at the border between the Columbia Plateaus and the Cascade Mountains and includes parts of these highly contrasted geologic provinces. Brief descriptions of the bedrock geology, together with geologic maps, may be consulted in other reports (Waters, 1932, 1938). For the purposes of this paper a much simplified geologic map (Pl. 1) has been prepared to show in particular the distribution of the major rock units with respect to their resistance to erosion as well as their relation to remnants of the "Entiat stage." By far the most resistant rocks of the area consist of a pre-Tertiary complex made up principally of the Chelan granodiorite and the Swakane gneiss into which the granodiorite is intrusive. These rocks are found in the Chelan and Entiat ranges and in the valley of the Entiat River. The least resistant are a group of arkoses, shales, and other varieties of sedimentary rock belonging to the Eocene Swauk formation. They are widespread in the broad synclinal valley of Wenatchee River, a small part

of which is shown in Plate 1. These two areas, underlain by rocks of widely different erosional susceptibility, are separated from the Waterville Plateau to the east by the sinuous canyon of the Columbia River. This plateau is underlain by a heterogeneous series of basalt flows, palagonite tuffs, pillow lavas, and associated sediments. The basalts of the plateau are a part of the widespread Miocene Columbia River basaltic series, but, due to the interbedding of the flows with sediments and the common condition of the volcanic material as pillow lava or as sideromelane and palagonite tuff and breccia, the series offers much less resistance to streams in this marginal part of the Columbia River lava field than do the massive uniform flows of the central part of the great lava plateau. For that matter, the resistance of even thick massive basalt flows varies considerably, depending upon the nature of the erosional agent. Although often cited in the literature as one of the most resistant rock types, massive basalt succumbs easily to the attack of large rivers because of the abundance of joints. Small streams, unable to pluck away the larger columns, encounter much more difficulty, and, if the region is not exceptionally well watered, most of the drainage disappears underground due to the high permeability of the well-jointed flows. Because of these factors the larger streams of the Columbia Plateau have formed impressive trough-like valleys along whose sides the smaller tributaries are left as "hanging valleys." It is also because of this factor that large rivers of glacial meltwater which crossed the plateau in Pleistocene time were able to accomplish prodigious feats of erosion; and yet the upland surface over large sections of the plateau has remained only slightly modified by normal erosional agencies since its formation near the end of the Miocene. Erosion here has been abnormally slow throughout post-Miocene time because of the great permeability of the lava flows.

Reference to the Chelan, Wenatchee, and Chiwaukum quadrangles of the Topographic Atlas of the United States reveals that topographically the three areas just described have little in common. The area underlain by the pre-Tertiary crystalline complex is characterized by gently undulating profiles of the "Entiat stage" which break off sharply to the Twisp canyons below. The relief is great, commonly 3000 to 5000 feet, but streams are widely spaced. In the area of Swauk sediments the uplands have been cut into by much more closely spaced canyons which do not show the striking "two storey" character of the valleys farther north. The drainage divides are sharpened to lines, and here and there ridges of harder arkose and conglomerate project as hogbacks above the general level. Numerous subsequent valleys have developed in the softer shales. The region presents the typical features of an area of moderately soft sedimentary rocks dissected to maturity. The contrast of these two areas

with that of the Waterville Plateau is even greater, for the surface of this plateau is an almost level plain diversified by a few basaltic mesas and by the shallow valleys of intermittent streams which locally have rather thoroughly dissected a thin series of sediments interstratified in the upper part of the basaltic flows to produce a series of rolling hills with an average relief of only 50 to 300 feet. The great contrast between this area and the mountainous district west of the Columbia impressed Willis (1903, p. 51) who wrote:

"There is in fact no separation in the sense that the mountain district west of the river has had an orogenic history different from that of the plateau east of it. But the rock structures and conditions of sculpture are so unlike that the two provinces appear at first sight to be physiographically and geologically distinct."

The first stage in this history as conceived by Willis was the reduction of all these diverse rocks to a common level to form the Methow Plain.

RESIDUALS OF THE METHOW PLAIN

"The characteristics by which areas of the Methow plain are identified are flatness, unsympathetic relation to environment, and relative elevation." (Willis, 1903, p. 71.)

Many residuals of the Methow Plain identified by Willis occur within the southern part of the Chelan quadrangle (Willis, 1903, Pl. 16). On the basalt plateau they are quite extensive, coinciding over large areas with the higher surfaces of this nearly flat plain. West of the Columbia they are smaller and few in number but are more striking physiographically because they are so completely out of harmony with the topography they surmount.

In the writer's opinion these residuals in the Chelan quadrangle are not remnants of a peneplain, but are surfaces determined by the structure of basalt flows without regard to a widespread base level of erosion. The areas east of Columbia River will be considered first.

Since Willis' report was written, road metal and sand pits have been opened in the "Two isolated hills southeast of Waterville" (Willis, 1903, p. 72; pl. 16) whose surfaces were regarded as remnants of the Methow Plain. These pits show clearly the structure of the two hills (Pl. 2, fig. 1). Their surface consists of a thin coating of loess resting upon a single basalt flow which covers stratified clays and sands about 200 feet thick. These sediments constitute the lower part of the hills but due to the mantle of loessial soil they are not visible except in recent cuttings. Wells and stream-cuttings near the bases of the hills show that the sediments in turn rest upon a thicker succession of basalt flows which contain local sedimentary intercalations as well as much pillow lava and palagonite. Clearly, the hills have the form of small mesas such as develop so typically

FIGURE 1. MESA SOUTHEAST OF WATERVILLE
Basalt flow caps the hill. This rests on stratified sands and clays which overlie basalt flows.

FIGURE 2. RIBBON MESA
Flat-topped mountain in center of the view is capped by an outlier of basalt which rests on granodiorite. Resurrected surfaces are visible to left and right of mesa surface.

FIGURE 1. LOOKING NORTHWEST
Camera was located at point B in Figure 2. Resurrected surface extends back from point A. Ribbon Mesa at extreme left.

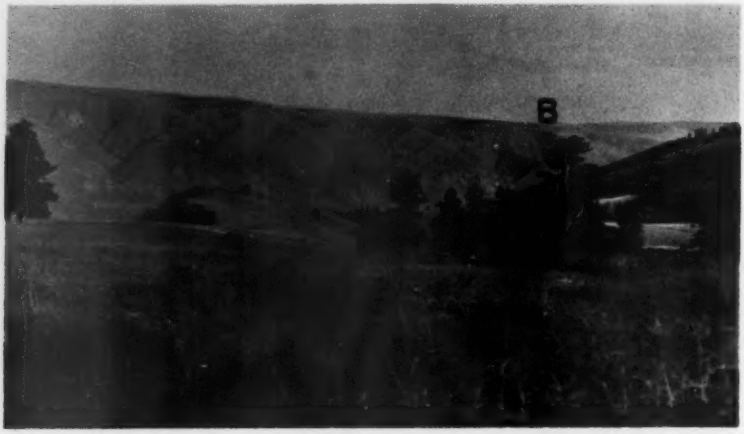

FIGURE 2. LOOKING SOUTHEAST
Camera was located at point A in Figure 1. East wall of Columbia Canyon rises abruptly to level of Waterville Plateau unbroken by a resurrected surface such as that forming the foreground.

WALLS OF COLUMBIA RIVER CANYON

in semi-arid regions where a thin hard stratum of horizontal material (in this case the basalt flow) caps much weaker material below.

The much more extensive flats near the brink of the Columbia Canyon in the northern and northwestern parts of the Waterville Plateau are of similar nature. Although the mantle of loess and fine-grained volcanic ash is very thick in these regions, road cuts and wells not available to the earlier investigators show that the sediments exposed in the Waterville mesas extend widely under this section of the plateau. In the areas where the supposed peneplain remnants occur these sediments are overlain by a thin capping of basalt whose structure corresponds exactly with the surface slope. These surfaces are commonly inclined at a very low angle, for Waterville occupies the center of a shallow spoon-shaped syncline in the basalt flows.

Over several square miles immediately north and west of Waterville the protecting basalt flow has been removed, and the intermittent streams have made appreciable progress in the soft sediments below. Since they are not powerful enough to pluck the basaltic columns in the thicker succession of flows underlying the sediments, the surface of these lower flows has formed a temporary base level of erosion. Consequently, the rolling hills carved from the sediments have a relief which does not exceed the thickness of the sedimentary intercalation.

Are the flat surfaces described of constructional or erosional origin? They may well represent remnants of the original surface of the basalt plain and so be surfaces of accumulation and not degradation. In this case they would date from the cessation of the basalt floods near the end of the Miocene, or perhaps in the lower Pliocene. But they may just as likely be erosion surfaces produced by the stripping of soft unconsolidated sediments from above the harder basaltic material. The essential point, however, is that such stripping can occur without reduction to a base level controlled by the sea. Stripping of the soft 200-foot thickness of sediment interstratified near the top of the basaltic series is going on today in the immediate vicinity of Waterville at elevations 2000 to 4000 feet above sea level. Over much of the summit and the northern slope of Badger Mountain this stripping has proceeded to completion, and a new mesa surface has appeared which is stratigraphically lower than that forming the top of the Waterville mesas but, due to warping actually exceeds the latter in elevation. Certainly none of these surfaces, coinciding as they do with the structure of the basalt sheets, can be taken as evidence of the general reduction of the entire region to a peneplain near sea level.

The smaller and more isolated residuals of the Methow Plain included in the Chelan quadrangle west of Columbia River are nearly all outliers

of the basaltic lavas of the Waterville Plateau separated from the main lava plain by erosion. Throughout this section of central Washington a thick veneer of loess and crystal ash of a pale tawny color mantles all flat surfaces and in places almost completely obscures the thin basalt capping, giving the deceptive appearance that the flats are cut across gneiss or granodiorite. This was confusing to the early workers since they considered the soil a residual product of the decay of the underlying rock (Russell, 1897, p. 58; Willis, 1903, p. 51-52). However, petrographic study of this soil shows that it is entirely of aeolian origin and in large part of volcanic derivation. Most of it is made up of grains of augite, hornblende, and andesine which commonly have films or shards of glass clinging to them, together with minor amounts of small pumice fragments. These materials have been derived from the andesitic volcanoes of the Cascade Range. The soil is not entirely of volcanic origin, but contains variable though often predominant amounts of mineral grains derived from plutonic and metamorphic rocks, some of which are not indigenous.

Careful search along the margins of the supposed peneplain remnants seldom fails to reveal one or more basalt flows or the associated pillow lavas, palagonite breccia, and Miocene sediments beneath this aeolian soil mantle. At Ribbon Mesa, the largest of the residuals, the soil mantle has been largely stripped away; consequently, the basalt is much in evidence and forms a columnated escarpment above the lower granodiorite slopes. The flats near the east end of the Entiat mountains clearly reveal the basalt on their southern margins, but toward the west and north it is much obscured by soil. At one locality in the Entiat Mountains no continuous remnant of a flow was found, but many large basalt fragments resting upon the gneiss reveal that a basalt outlier which once covered this area has now been disintegrated and largely removed. On the large flat marked as a residual of the Methow Plain located at the crest of the Entiat Mountains near the western margin of the Chelan quadrangle the writer found no basalt. The foliation of the gneiss is horizontal at this locality, and two possible explanations of this flat other than that it is a peneplain remnant are: (1) It may once have been covered with basalt, or more likely with the sediments associated with the basalt, and these have been stripped away; (2) it may be merely a structural platform carved from the horizontally foliated gneiss.

Wherever the exposures are sufficiently continuous to make out the structure the inclination of the surface and of the basalt flows corresponds. This is well shown at Ribbon Mesa and at the residuals on the Waterville Plateau. Willis found that one of the flats near the

eastern end of the Entiat Mountains was underlain by basalt, and concerning its evidence on peneplanation he wrote (Willis, 1903, p. 64):

> "The broad flat at the eastern end of the Entiat Range, 4,500 to 4,900 feet above sea, corresponds in extent with a basalt remnant. The dip of the basalt is not less than 3 degrees and probably nearer 5 degrees to the north. The slope of the surface is about 1½ degrees in the same direction. The dip and slope therefore appear to diverge 150 feet or more to the mile. If we assume that at any time the surface was level, the basalt then dipped 1½ degrees, and the level crossed the basalt beds at a slight angle. It is more probable that the surface declined slightly down the course of the Columbia and was cut across the nearly level basalt. Both have been tilted in the uplift of the Entiat Range.
>
> "Though small, the angle at which the flat surface and the basalt bedding diverge is important, since the surface is thereby identified as one of erosion in contradistinction to one of structure."

The writer can neither confirm nor deny Willis' observation. The basalt appears to consist of a single massive flow. It is not well exposed, and in most places only small areas of irregular columns are visible. The presence of scoriaceous material directly below the aeolian soil at four different localities on the outlier lead the writer to conclude that flow and surface correspond and that both are nearly horizontal; but it was impossible to get a dip reading on the flow which would be accurate within 5 degrees. Granting a divergence of 1½ degrees between the dip of the flow and the slope of the surface, such a difference over an area of less than a square mile would not necessitate that the erosion surface be produced as part of a peneplain near sea level.

In his report, Willis (1903, p. 64) clearly indicated that "The evidence is slight and by itself insufficient to establish a conclusion."

METHOW OR CASCADE PLAIN IN THE YAKIMA DISTRICT

A less dubious example of erosional bevelling of basalt flows has been described by Smith from the Yakima district at a point east of Kelly Hollow in the drainage area of Wenas Creek. This locality is approximately 75 miles south of the Wenatchee-Chelan district, but inasmuch as the conditions of structure and climate are much the same in the two areas, and since the Kelly Hollow locality is often referred to as the "type locality" for the Methow or Cascade Peneplain, some discussion of it in this report may be warranted.

At this locality there is no doubt that the basalt flows and the associated Ellensburg sedimentary rocks have been truncated to a common surface by erosional processes. The writer can confirm in every detail the description and sketch given by Smith (1903a, p. 27). The relationship is the more striking because of the complete effacement of the contact between the hard basalt and the soft Ellensburg sediments by the new surface. But one characteristic of this surface seems to rule out its interpretation

as the remnant of a widespread peneplain formed near sea level. This characteristic was noted by Smith (1903a, p. 27) who wrote: "The surface is thickly strewn with angular bowlders of basalt." Many of these boulders are more than 3 feet long. They are scattered as thickly over the area underlain by Ellensburg sediments as they are over the truncated basalt flows. These angular blocks are surprisingly unweathered. Such materials seem to be quite out of place as a veneer on a peneplain surface developed near sea level. Instead, they resemble the waste scattered over the surface of desert plains as a result of the action of sporadic "cloudburst" floods and mudflows. The surface at Kelly Hollow is interpreted by the writer as a dissected pediment of relatively small extent. Buwalda (1936) has suggested that it is a graded surface developed by tributaries of Wenas Creek when this creek flowed at a higher level. As Smith (1903a, p. 33-34) points out, the Wenas Creek drainage has undergone several complicated adjustments as a result of recent warping; consequently, surfaces once graded are now out of harmony with their surroundings.

In general, relations such as those found at Kelly Hollow are rare in the Yakima district. Most of the prominent anticlinal ridges (Smith, 1903b) formed from the basalt flows reveal a very close correspondence between surface and inclination of the flows.

However, at a few places on the Columbia River Plateau graded surfaces which truncate basalt flows at an angle are developing today along the margins of intermontane basins at various levels above that of the sea. One such surface in process of development can be studied at the northern margin of the Quincy Basin where several flows are being truncated and levelled by processes of desert erosion. Similar relations may be seen locally at the foot of the Grand Coulee monocline and along the northern base of Badger Mountain. All these graded plains are developing high above sea level.

In north-central Oregon an erosional plain truncating the basalt flows has been observed in the Ochoco region and has been described by Buwalda (1929) as the Ochoco Peneplain. That the Ochoco "Peneplain" is the result of widespread reduction of the surface to a level approaching that of the sea may be questioned, however, in view of the fact that Gilluly, Reed, and Park (1933, p. 24) were unable to find evidence of extension of this peneplain in their studies of the immediately surrounding areas in the Blue Mountains.

RESIDUALS OF THE "ENTIAT STAGE"

"The characteristic topographic form of the Entiat stage is mature. It occurs as a spur or divide below occasional residuals of the Methow stage and above features of later stages." (Willis, 1903, p. 68.)

Some of the best examples of Entiat stage surfaces are found in the area southwest of Chelan Butte (Pl. 4, fig. 1) and along the higher slopes bordering the lower course of Entiat River. The Entiat residuals are undoubtedly true surfaces of erosion and clearly represent remnants of a once-continuous mature surface which formed a mountainous upland with a relief of 1500 to 2500 feet.

A comparison of the distribution of the Entiat stage residuals with that of the various rock formations in the area reveals certain relationships. The Entiat surfaces are found only on the most resistant rock types of the area. They are best developed on the massive granodiorite of the Chelan batholith but are also conspicuous in areas underlain by the Swakane gneiss. In areas of phyllite and well-foliated schist they are very poorly preserved and in the area underlain by the Swauk formation they cannot be recognized with any certainty. But the most striking feature of their distribution is their complete absence from the areas underlain by the Columbia River lavas.[2] This relationship is rendered all the more peculiar because the best-preserved Entiat surfaces occur along the western margin of the Columbia River valley just across the river from the lava plain of the Waterville Plateau. Here, in the area between Chelan Butte and the Entiat River, they form long sloping spurs that sweep down from the higher summits of the mountains to levels 800 to 1500 feet above the Columbia where they break off to the steep-sided Twisp stage canyon. However, when one looks for the continuation of these surfaces on the eastern wall of the canyon one sees only basalt escarpments which rise abruptly to the surface of the Waterville Plateau with no trace of the long sweeping spurs.

Another curious fact about the distribution of the Entiat residuals is that, although they are not developed across the basalt itself, they are always best developed where the granodiorite or gneiss of the old pre-Tertiary complex comes in closest proximity to the basaltic escarpment of the Waterville Plateau. Thus, superlative examples occur at the eastern end of the Chelan Mountains and along the lower course of the Entiat River, but, when traced farther to the west, these surfaces become more and more fragmentary and indefinite until in that part of the Chelan Mountains lying 20 to 25 miles northwest of the border of the lava field they can be recognized only with difficulty.

The absence of Entiat features from areas underlain by the Columbia River basalt is not unique to the Wenatchee-Chelan district. In the

[2] On the limbs of some of the anticlinal ridges (Smith, 1903b) of the basalt plateau small streams flowing directly down the limb of the fold may show an upper and wider valley above a younger narrow canyon. The composite character of these valleys is due to the acceleration of the streams as a result of anticlinal rise and not to a change of baselevel by vertical uplift of the entire region. Such valleys, restricted to anticlinal limbs, are not to be confused with features of the Entiat stage.

Yakima area where the surface rocks consist almost entirely of these lavas and the associated Miocene sediments, Smith (1903 a, p. 30-31) found no physiographic features which he could correlate with Willis' Entiat stage.

The absence of Entiat stage features from the basaltic plateau is explained by Willis (1903, p. 79) as follows:

> "In arid areas Entiat features have not been clearly recognized. This fact may be due to one of two conditions. In Entiat time areas now arid and high may have been similarly situated, or they may have been low. In the one case, aridity, in the other lowness, would have limited erosion, and the features which developed would in either case be inconspicuous. In the Yakima district G. O. Smith does not distinguish the Entiat stage. On the Waterville plateau and Badger Mountain it is probable that Entiat features exist, but are mantled by later deposits and obscured."

The possibility of the existence of Entiat Stage valleys filled with later debris in the Waterville Plateau-Badger Mountain area can now be disregarded as these areas have been geologically mapped and their subsurface structure is well known. Difference in aridity can hardly explain the absence of Entiat features on the Waterville Plateau, for this district is separated only by the canyon of Columbia River from the equally arid ridges of the eastern end of the Chelan Mountains where the most perfectly preserved Entiat surfaces are found. Nor can a difference in prior elevation be the explanation, for a warp or other structural disturbance responsible for such a difference in elevation would have to follow the sinuous course of Columbia River to produce the desired result. Furthermore, such warping is excluded from consideration by the occurrence of outliers of the Columbia River basalt on the west side of the river at levels appropriate to those on the Waterville Plateau.

The hypothesis adopted in this paper to explain the distribution of Entiat surfaces with relation to the various kinds of bedrock is that these surfaces represent areas once buried beneath the Columbia River basalts and their associated sediments and then resurrected by the erosion of the Columbia River and its tributaries since the cessation of the basaltic floods. Before developing this hypothesis in detail, let us consider the stratigraphy and manner of accumulation of the Columbia River lavas in this part of the plateau.

NATURE OF SURFACE BURIED BY THE COLUMBIA RIVER LAVAS

Brief accounts of conditions prevailing when the lavas of the Waterville-Moses Coulee area were accumulating have been published by Fuller (1931) and by Chappell (1936). The lavas were extruded from fissures located to the east and southeast of this area. The lava flows, spreading to the north and west, advanced on a topography of considerable relief. By tracing the contact of the basalt with the basement rocks along

the margin of the Waterville Plateau, it can be seen that the flows have buried hills and valleys having differential elevations of as much as 2100 feet. The drainage of this mountainous upland was toward the center of what is now the Columbia Plateau; consequently the lava floods, advancing up the former stream valleys, blocked the drainage and formed a series of marginal lakes at the border of the encroaching lava field. In these shallow lakes the streams spread out deltaic accumulations of sand and clay brought down from the adjacent highlands, and when there were long periods of halt between successive flows these sedimentary accumulations were able to spread some distance toward the interior of the lava plain. More often, an individual marginal lake would last for only a short time before it and its accumulated sedimentary deposits were overwhelmed by a new flow of lava advancing from the interior of the plateau. Such a flow, upon entering the shallow body of water, would granulate into slag-like masses of basaltic glass now commonly altered to palagonite or, if the magma was more viscous, would form typical pillow lava (Fuller, 1931). These pillow lavas and palagonite breccias generally ended distally against the deltaic accumulations of sediment in the former lake. Only rarely would a flow reach and spread out over the borders of the old granitic upland without encountering these marginal bodies of water. Consequently, much of the volcanic material occurring immediately at the margin of the lava field is not in massive flows but is in the form of pillow lavas or palagonite tuffs interstratified with sediments. With each advancing flow the position of the marginal lakes was shifted farther and farther to the west. Therefore, there is seldom a great thickness of sediment or volcanic material showing the effects of aqueous chilling at any one place, and these aqueous accumulations are commonly covered with a considerable thickness of basalt which shows normal subaerial characteristics.

The important point for the present discussion, however, is that there occurs along the margin of the basaltic series in this part of central Washington a heterogeneous zone of variable thickness where much palagonite, pillow lava, and more or less poorly consolidated sedimentary material are found in the volcanic succession. These materials are very easily removed by erosion, and even massive basalt flows which lie above them are easily removed by the sapping of the underlying softer deposits and resultant landsliding.

At the cessation of the basaltic floods the low area of marginal lakes determined the position of the present Columbia River, which from its junction with the Spokane to a point a few miles below Wenatchee follows very closely the northwestern margin of the basaltic plain. As the river began to incise its present canyon, the soft and easily eroded sedimentary

and palagonitic accumulations would naturally be the first materials removed. Where these materials rested upon the very resistant Chelan granodiorite and Swakane gneiss the old mature topography buried below the volcanic succession was quickly exhumed; but, where they rested upon the soft Swauk formation, differential stripping did not occur. The Entiat surfaces seen so clearly upon the crystalline rocks forming the west wall of the Columbia Canyon are resurrected surfaces. Once buried beneath the sediments and volcanic accumulations of Miocene age, they have been etched out on account of the great difference in resistance to erosion between the underlying granodiorite and its volcanic cover.

LOCAL DETAILS OF THE PROCESS OF RESURRECTION

The process of stripping of the volcanic cover is best illustrated in the area just north of Ribbon Mesa (Pl. 1). Here the stripping appears to be of comparatively recent date. Consequently in the area near the basalt outlier that forms the top of the mesa, loose fragments of basalt and occasional basalt remnants frozen to the surface of the underlying granodiorite survive as evidence of the former continuity of the volcanic cover. Traced upward, the Entiat slopes disappear against the margin of the basalt outlier. Where the exposures are good along the base of the basalt columns that form the mesa rim, it can be seen that the lava flowed out upon an irregular surface which in its form is the direct continuation of the surface now exposed as Entiat stage slopes lower down on the mountain.

Some distance north of Ribbon Mesa, practically all traces of the original volcanic cover have disappeared, but at two localities small isolated masses of palagonite breccia, neither more than a few square yards in area, may be seen plastered upon the exhumed Entiat slopes. A comparison of the two walls of the canyon of Columbia River at this locality is very instructive (Pl. 3). On the west side of the river, sharp slopes belonging to the Twisp stage rise abruptly from the water's edge to heights of 1700 feet. From these the more gentle slopes of the Entiat stage, mantled in most places by a coating of the prevalent aeolian soil, sweep up to the ridge which forms the divide between the Columbia and the Entiat drainage. Farther south, they rise and disappear beneath the basaltic outlier of Ribbon Mesa. Looking across Columbia River to this part of the Chelan Mountains from a high vantage point at the brink of the Waterville Plateau, the observer finds that these rolling Entiat surfaces are the most conspicuous things in the view (Pl. 3, fig. 1). If, however, he reverses his position and stands on one of the Entiat stage spurs west of the river and looks eastward to the Waterville Plateau (Pl. 3, fig. 2), the view is strikingly different. The steep lower slopes

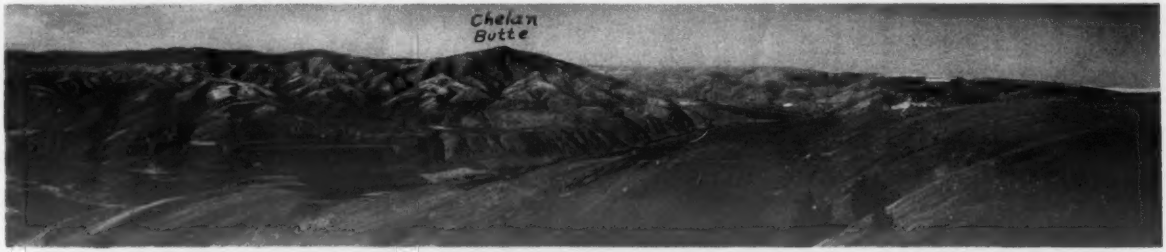

FIGURE 1. RESURRECTED SURFACES NEAR CHELAN BUTTE
Basalt plateau is shown at right. Resurrected surfaces (many of them cultivated) form the ridgetops to left of Chelan Butte. Camera was located at point B of Figure 2.

FIGURE 2. SOUTH WALL OF COLUMBIA RIVER CANYON OPPOSITE CHELAN BUTTE
Camera was located across the Columbia River on a resurrected surface at point A in Figure 1.

WALLS OF COLUMBIA RIVER CANYON NEAR CHELAN BUTTE

of the Columbia Canyon are there, though their abruptness is modified by a younger fill of stratified glacial drift. Above the level of the Twisp stage canyon, instead of the long sweeping profiles of the Entiat stage, horizontal basalt flows separated by steep grassy slopes rise in tiers completely to the level of the Waterville Plateau whose surface is not sculptured into a rolling matureland of great relief like that of the mountains west of the river but is almost flat. However, close examination of the east wall of the canyon at approximately the level where the Entiat slopes should appear reveals an irregular line which marks the junction between the light-colored granodiorite below and the black basalt above. This line is not horizontal but rises and falls when traced along the wall of the canyon, approaching in some places within 150 feet of the plateau surface, sinking in others to levels more than 2000 feet below it. Furthermore, in a few localities erosion has begun to etch out this surface of contact and locally has formed small terrace-like areas where the basalt has been stripped from the basement below. Nevertheless such areas of resurrected topography are very insignificant in extent compared with those on the west side of the river.

Similar relations can be observed farther upstream. Remnants of the Entiat surface are particularly well developed on the ridge which separates the Columbia drainage from that of lower Lake Chelan (Willis, 1903, p. 77). In a view of the region they are conspicuous because, though steep, they are flat enough and sufficiently mantled with aeolian soil to support wheat farms, whereas the Twisp ravines cutting into them are too steep sided for agricultural pursuits. Just west of Chelan Butte, long undulating Entiat spurs (Pl. 4) sweep down from the crest of the ridge, 3000 feet above sea level, to a point 2½ miles farther south, where, at an elevation of 1500 feet, they break off abruptly to the level of the Columbia River about 800 feet below. No similar slopes break the abruptness of the opposite canyon wall (Pl. 4, fig. 2), but at approximately the spot where a continuation of these features should appear a small basaltic remnant caps a hill about 1100 feet above river level. A short distance back of this point the dissected ends of basalt flows appear in an escarpment that rises abruptly to the level of the Waterville Plateau 2400 feet above the river (3200 feet above the sea). Immediately to the west of this point a short deep ravine has trenched the basalt and reveals the fact that the surface which the volcanic rocks covered also sloped upward to the southeast at an angle which is only a little steeper than the Entiat slopes on the opposite side of the river. At the head of the gulch the basalt covering the granodiorite is only 300 feet thick.

Why are these resurrected surfaces so much more conspicuous on the west side of the Columbia than on the east side? Several factors seem

to be involved. Most important perhaps is the fact that the volcanic succession thickens to the east; consequently, the farther east one goes the more material there is to be removed. This thickening is rather abrupt. From less than 150 feet of volcanic material at Ribbon Mesa the succession thickens to about 700 feet 6 miles to the southeast, and where the Badger Mountain anticline crosses Douglas Creek another 18 miles to the southeast 1800 feet of volcanic material is found with the base not exposed.

Another important factor is that the western or marginal portions of the volcanic succession contain much pillow lava, palagonite breccia, and poorly consolidated sedimentary deposits which are easily removed.

Another factor, perhaps most important of all, is the fact that the courses of the tributary streams west of the Columbia were in all probability determined before the arrival of the basalt floods, and these rivers, such as the Entiat and the Wenatchee, probably flow today in valleys not far removed in position from the valleys they occupied on the old Entiat matureland of Miocene time. The Entiat surfaces were therefore adjusted in slope and form to the position of the present valleys, and where the old valleys were filled with volcanic deposits the streams have extended as far as the Columbia along approximately the same lines. The present Entiat River, for example, is bordered by exhumed Entiat surfaces along the lower part of its course that would be in adjustment with a stream occupying almost the same position at a higher level. This adjustment of direction of buried slope to present stream positions would aid in the rapidity with which the surface would be resurrected. On the other hand, the Columbia River and those tributaries which join it from the east are entirely out of harmony with the pre-lava drainage system since they have developed as consequents on the surface of the new volcanic plain. Where they have cut through the volcanic cover into the older land surface below, they are out of adjustment with the buried topography, and therefore resurrection of the pre-volcanic stream pattern by stripping is much slower. The Columbia in particular is decidedly out of adjustment and appears to flow practically at right angles to the previous drainage lines. As a result, good cross-profiles of pre-volcanic valleys and ridges are commonly revealed by the contact line between the granodiorite and the basalt on its eastern wall.

At a few places areas of resurrected surface of considerable extent have developed on the east side of the Columbia. One such area is in the drainage area of Corbaley Canyon near the northwestern corner of Badger Mountain. Here the Pliocene and Pleistocene warping responsible for the rise of the mountain as an anticlinal ridge have tilted the buried

surface at the base of the volcanics toward the axis of Corbaley Creek, greatly facilitating stripping.

An interesting resurrected surface is found as a narrow terrace which slopes southward on the eastern wall of the Columbia Canyon at the southern margin of the Chelan quadrangle. This is probably resurrected from the base of the basalt only at its northern end. Farther south it appears to be a surface exhumed from beneath the soft shales and thin conglomerates at the base of the Swauk formation. These beds rest on the more resistant Swakane gneiss. Unlike the Entiat resurrected surface it is not a hilly matureland but is flat and featureless.

APPLICATION TO OKANOGAN REGION AND HIGHER PARTS OF CASCADE RANGE

Dawson (1891), Umpleby (1910, 1912), and other geologists have described a peneplain surface developed across the crystalline rocks of the Okanogan Highlands and the northern part of the Cascade Mountains near the Canadian boundary. Daly (1913), on the other hand, found no evidence of such a surface and, in criticism of the results of Smith and Willis, he pointed out that several factors operative in mountainous regions may bring about a striking accordance of summit levels. Others (Blackwelder, 1912; Rich, 1918), though accepting the existence of the peneplain described by Umpleby, have criticized the dating of it as Eocene.

The writer claims no very detailed geologic knowledge of the more rugged parts of the Cascade Mountains west of the Okanogan Range, but after numerous trips into this region he is inclined to agree with Daly that there is no evidence of a widespread peneplain in this area. Even the accordance of crest lines, so often described as exceptionally striking, seems more apparent than real. In the clear mountain air of this district the form of a mountain slope even 40 or 50 miles away can be made out distinctly. At such distances mountain peaks with differences in elevation of thousands of feet integrate readily into what appears to be a fairly continuous skyline because of the haphazard spacing of peaks of different heights. Even on such a skyline in the northern Cascades, however, many conspicuous mountain summits can be seen which tower to varying heights above their fellows. Profiles plotted for those parts of the area that have adequate base maps show no one level to which the peaks approach but emphasize, instead, their great variation in height. Practically all the peaks and ridges have sharp crests; many are pinnacled glacial horns.

However, while engaged in mapping the geology of the Chopaka and Osoyoos quadrangles which lie at the junction of the Cascade Mountains

and the Okanogan Highlands immediately east of the extremely serrate region just described, the writer found at many localities rounded or flattish crests on the mountain summits which appear to be remnants of an old erosion surface. These are apparently the surfaces described by Umpleby as remnants of a peneplain which he believed extended not only over this area but widely to the west and north and merged on the east with the much-discussed peneplain surface of Idaho. It is also considered to be the southern continuation of the erosion surface found in the Interior Plateaus of British Columbia.

Whatever the condition of the surface in Idaho and adjoining regions, the writer does not believe that in the Okanogan and Colville ranges it should be termed a peneplain. Though obviously an erosion surface, it was far from flat but appears to have had an undulating mature surface with differential elevations locally as much as 3000 feet. Today it exists in a decidedly fragmentary condition and in some parts of the area cannot be recognized with any certainty.

It is fairly well developed in parts of the Okanogan Range. Here the underlying rock is the massive granodiorite of the Similkameen batholith. The high, broad summits of this range are incised with old valleys cut into the granodiorite and in places partially filled with younger deposits, chiefly of pyroclastic origin, which are decidedly out of harmony with the sharply incised canyons that carry the present drainage of the range into the Similkameen, Methow, and Okanogan rivers. Similar surfaces are somewhat better developed on the Colville batholith east of the Okanogan River. Here many broad, high-level valleys are present which are quite out of harmony with their surroundings. For example, the upper courses of Chewiliken, Bonaparte, and Tunk creeks (Osoyoos topographic sheet) are broad and flat bottomed, but upon leaving them the streams descend abruptly to the much lower Okanogan River in narrow trenches with numerous cascades. These high-level surfaces become even better defined as one approaches the great basalt outlier which lies north of the Columbia River in the Colville Indian Reservation. In the Sanpoil and Nespelem valleys east of this outlier, Pardee (1918) recognized a similar high-level surface which he named the Sanpoil erosion surface. Concerning its relation to the Columbia River basalt, he wrote (p. 45):

> "The Sanpoil erosion surface is not to be confused with the surface represented by the Columbia Plain, the Okanogan Plateau, and the (basalt-capped) table-lands of the Nespelem Valley. Comparison of the two brings out the fact that the Sanpoil surface has suffered an incomparably greater erosion. . . . Moreover, . . . it appears that the Columbia Plain is a constructional surface and the lavas that built it are observed to lie upon the surface represented by the generalized slopes of the Nespelem basin. In other words, the deformed and eroded Sanpoil surface passes beneath the Columbia Plain."

In contrast to these areas, the district which extends west of the Okanogan River as far as Toats Coulee is a belt of rugged somewhat lower ground in which the old surface is very poorly preserved. This area separates the Similkameen and Colville batholiths and is underlain by Early Tertiary lavas and sediments and by a very heterogeneous group of phyllites, schists, greenstones, and marbles known as the "Anarchist series." On these rocks the mature erosion surface is rarely preserved. It can be recognized with a fair degree of certainty on Elemeham and Kruger mountains, less clearly in the so-called "Lime Belt" northwest of Riverside, and doubtfully if at all on Whisky Mountain and the other mountains of the district.

This brief description indicates that this surface in the Okanogan region has many of the peculiarities which characterized the Entiat surface in the Wenatchee-Chelan district 80 miles to the south. It shows the same avoidance of areas of soft phyllite, chlorite schist, or Tertiary sedimentary rocks and is well developed only on massive crystalline rocks. As in the Entiat region, the perfection of the residuals increases as the margin of the Columbia River basalt is approached; and like the Entiat it is a mature surface, not a peneplain.

Is this surface, like that of the Entiat matureland, an old landscape resurrected from beneath the Columbia River lavas? This explanation might seemingly apply to the district immediately adjacent to and east of the basalt outlier of the Colville Indian Reservation as Pardee (1918, p. 45) has already demonstrated, though the evidence is not as striking as in the Chelan district. For the Okanogan Range and the northern part of the Colville mountain group, however, this hypothesis certainly cannot be rigidly applied for there is no evidence that the basalt ever extended as far north and west as these localities. Nevertheless when traced to the south and east, these surfaces appear to coincide with the surface which disappears under the lavas. Also in the Wenatchee-Chelan district fragmentary portions of a surface which is clearly a continuation of the Entiat surface are found in the western part of the Chelan Mountains in areas that were almost certainly not covered by the basalt. Is it then possible that in central and northern Washington a mature erosion surface developed in Miocene time and was then buried by the basalt floods to the east but in the western part of the area has remained exposed to the agents of erosion since the cessation of the basaltic outpourings (probably late Miocene or early Pliocene) and is still recognizable in the landscape? Considering the rapidity with which erosion is believed to have taken place in our western mountains during the late Tertiary, the possibilities of survival of such a surface even on the hardest and most resistent rocks appear none too favorable, but two rather peculiar cir-

cumstances in this particular area have tended to preserve in a more or less unmodified condition surfaces that date back to approximately the end of the Miocene. These are (1) the ponding and disruption of the ancient drainage system by the outpouring of the Columbia River basalt and other volcanic series, and (2) the burial of the surfaces time after time by showers of pyroclastic material as a result of the recurrent explosive activity of the andesitic volcanoes in the Cascade Mountains to the west.

The effects produced immediately at the margin of the lava field as a result of ponding of the ancestral drainage by the Columbia River flows have already been considered. The effects farther back within the main mass of the highlands probably would not be great, but the lakes developed so conspicuously at the lava margin would doubtless extend some distance up the larger graded valleys and would become traps for sediment. The general effect would be to slow down the rate of erosion as a result of the substitution of a higher base level and because of the temporary detention of a part of the detritus in inland lakes. However, in regions which stood higher than the surface of the basalt plain these effects would doubtless be unimportant or nonexistent.

A far more effective means of slowing the rate of degradation is to be found in the showers of pyroclastic material which almost certainly covered this area time after time throughout the Pliocene and Pleistocene. Even today the Entiat surfaces in the western parts of the Chelan Mountains are deeply buried under an accumulation of pumice lapilli blown out from the Cascade volcanoes in postglacial time. This pumice blanket, 2 to 10 feet thick in the upper part of the Entiat River drainage, thins to the east and passes into a thin layer of white ash. Similar volcanic products deeply mantle the flatter parts of the old erosion surface across the Okanogan Range. These late Quaternary deposits are doubtlessly insignificant compared with the tremendous volume of pyroclastic debris that must have settled over this area during the much more pronounced volcanic activity of the Pleistocene, and it is probable that even the Pleistocene accumulations were dwarfed by those associated with the great volcanic activity responsible for the building up of the widespread Keechelus andesitic series of late Miocene and Pliocene age. The presence of a thick cover of pyroclastic debris on every surface that is flat enough to hold it is one of the conspicuous features of Cascade geology.

The effect of these repeated showers of pyroclastic debris in slowing erosion in certain areas may have been marked. On steep slopes they would have relatively little effect, for, due to its very low specific gravity, the material is easily removed by running water, but on flat surfaces

the problem would be very different. Here the important factor to consider is the extremely high permeability of the deposits. Rain falling on them would not collect in rivulets but would immediately sink underground. It is well known that certain areas covered by pumiceous lapilli and ash in the Oregon Cascades near Crater Lake are devoid of streams and support little vegetation—the few plants found are chiefly of a desert type—despite the fact that the area receives an extremely heavy rainfall. The main effect of such a deposit in slowing erosion is not due to the added material to be removed but to the fact that this highly permeable material acts as a sponge which sops up the rainfall and prevents it from concentrating in surface streams. This factor more than any other accounts for the fairly good preserval of portions of an erosion surface in the Okanogan Range and in the western part of the Chelan and Entiat Mountains. Probably the areas in the Methow Mountains mapped by Willis as remnants of the Entiat and Methow stages are of a similar nature.

The survival of the ancient surface thus depends upon a peculiar combination of favorable conditions. First, it had to be developed on hard rock which had been carved into very broad valleys and ridges with comparatively gentle slopes. In areas of softer material the more closely spaced ridges and valleys with intervening shorter but perhaps steeper slopes did not furnish a stable enough basement for the pyroclastic deposits to stick for any great length of time. When they were removed the soft rock below was quickly gutted, thus increasing the surface irregularity and rendering the next pumice blanket less effective as a protective medium. Still farther west, in what now constitutes the highest parts of the Cascades, conditions were also unfavorable for preserval due to the greater elevation with consequent sharpening of the slopes. If the former rainfall distribution was like that at present the much greater precipitation to the west would also be an important factor. Probably the pyroclastic accumulations were of considerably greater volume here than farther east, but falling on the steep slopes of a well-watered rapidly rising young mountain mass they were quickly swept away. Only farther east near the margin of the great lava plain where the land was lower, flatter, and drier did the blotter-like action of the pyroclastic material become really effective.

CONCLUSIONS

Thus, in proceeding from the margin of the lava plateau of central Washington northwest into the serrate mountains of the northern Cascades, one encounters first a well-preserved mature erosion surface which has been resurrected from beneath the Columbia River basaltic series

and its associated sediments. Beyond this zone the surface continues in a more fragmentary condition. Here it owes its preservation to burial under repeated showers of pyroclastic material whose blotter action stemmed erosion except on the steeper slopes. Traced still farther west into the higher parts of the range it gradually becomes less and less well defined as the result of increasing dissection until ultimately it is lost completely in the sea of ragged peaks that form the higher parts of the Cascade Chain.

Can this new hypothesis of the origin of the surface in the Okanogan Mountains and its correlation with the similar Entiat surface farther south furnish any information regarding the much-discussed problem of the age of the surface in the Okanogan district? The writer finds no evidence favoring Umpleby's conclusion that this surface is of Eocene age. In the Okanogan valley sediments containing plant remains of Early Tertiary age are truncated by what appears to be a remnant of this surface in an area east of Elemeham Mountain. These sediments are considerably deformed and stand vertical near Oroville. Similar sediments in the vicinity of Tonasket are also folded and show dips as high as 25 degrees. No such folding of the surface itself has been observed, though it is broadly warped locally. Since the surface disappears beneath the Columbia River lavas it must have attained its form as a matureland just prior to the outpouring of the basalts. The age of the beginning of lava extrusion is not closely defined despite an abundance of paleobotanical remains but it probably is Middle Miocene and certainly is not younger than Upper Miocene. Nevertheless, although the surface was given its form in Miocene time it does not follow that the landscape which it forms today is of Miocene age. Immediately adjacent to the Columbia River this Miocene surface was buried under hundreds, and in a few places even thousands, of feet of volcanic debris. The date of its resurrection to become an element in the present topography probably varies greatly from place to place. At one locality it may be Pliocene, at another Quaternary. Farther west beyond the reach of the basaltic floods portions of this surface are probably visible in the topography today which have never been covered very deeply under pyroclastic debris and which in the entire interval of time from the Miocene to the present may have undergone only minor changes in form as elements of the landscape. There seems to be no way of distinguishing such relatively static portions of the surface from others that have been buried once or perhaps many times to considerable depths beneath andesitic pyroclastics. Thus we are left with the rather interesting conclusion that the age of the surface as we see it today may be Miocene in one place, Pliocene in another, and Quaternary in a third. It is the product of a complicated series of

events that began with the outpouring of the Columbia River basalt and have continued to the present time.

WORKS TO WHICH REFERENCE IS MADE

Blackwelder, Eliot (1912) *The old erosion surface in Idaho, a criticism,* Jour. Geol., vol. 20, p. 410-414.

Buwalda, J. P. (1929) *A Neocene erosion surface in central Oregon,* Carnegie Inst. Washington, Publ. no. 404, p. 1-10.

——— (1936) *Postulated peneplanation in central Washington* (abstract), Geol. Soc. Am., Pr. 1935, p. 331-332.

Chappell, W. M. (1936) *The effect of Miocene lavas on the course of Columbia River in central Washington,* Jour. Geol., vol. 44, p. 379-386.

Daly, R. A. (1913) *Geology of the North American Cordillera at the forty-ninth parallel,* Geol. Survey Canada, Mem. 38, p. 1-799.

Dawson, G. M. (1891) *On the later physiographical geology of the Rocky Mountain region in Canada, with special reference to changes in elevation and to the history of the glacial period,* Royal Soc. Canada, Pr. Tr. 8, p. 3-74.

Fuller, R. E. (1931) *The aqueous chilling of basaltic lava on the Columbia River Plateau,* Am. Jour. Sci., 5th ser., vol. 21, p. 281-300.

Gilluly, J., Reed, J. C., and Park, C. F. (1933) *Some mining districts of eastern Oregon,* U. S. Geol. Survey, Bull. 846-A, p. 1-140.

Pardee, J. T. (1918) *Geology and mineral deposits of the Colville Indian Reservation, Washington,* U. S. Geol. Survey, Bull. 677, p. 1-186.

Rich, J. L. (1918) *An old erosion surface in Idaho; is it Eocene?* Econ. Geol., vol. 13, p. 120-136.

Russell, I. C. (1897) *A reconnaissance in southeastern Washington,* U. S. Geol. Survey, W. S. Paper 4, p. 1-96.

——— (1900) *Geology of the Cascade Mountains in northern Washington,* U. S. Geol. Survey, Ann. Rept. 20, pt. 2, p. 83-210.

Smith, G. O. (1903a) *Geology and physiography of central Washington,* U. S. Geol. Survey, Prof. Paper 19, p. 9-39.

——— (1903b) *Anticlinal mountain ridges in central Washington,* Jour. Geol., vol. 11, p. 166-177.

Umpleby, J. B. (1910) *Geology and ore deposits of Republic Mining District, Washington,* Wash. Geol. Survey, Bull. 1, p. 1-67.

——— (1912) *An old erosion surface in Idaho,* Jour. Geol., vol. 20, p. 139-147.

Waters, A. C. (1932) *A petrologic and structural study of the Swakane gneiss, Entiat Mountains, Washington,* Jour. Geol., vol. 40, p. 604-633.

——— (1938) *Petrology of the contact breccias of the Chelan batholith,* Geol. Soc. Am., Bull., vol. 49, p. 763-794.

Willis, Bailey (1903) *Physiography and deformation of the Wenatchee-Chelan district, Cascade Range,* U. S. Geol. Survey, Prof. Paper 19, p. 49-97.

——— (1936) *East African plateaus and rift valleys,* Carnegie Inst. Washington, Publ. 470, p. 113-128.

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PRESENTED BEFORE A JOINT MEETING OF THE CORDILLERAN SECTION OF THE GEOLOGICAL SOCIETY AND THE PACIFIC DIVISION OF THE AMERICAN ASSOCIATION FOR THE ADVANCEMENT OF SCIENCE, SEATTLE, JUNE 18, 1936, UNDER THE TITLE, *Erosion surfaces in the Wenatchee-Chelan District, Washington.*

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DISCUSSION BY BAILEY WILLIS

My colleague, Professor Waters, and I have discussed the physiographic development of the Lake Chelan-Columbia region in the field together and in the office. His thorough and well-balanced investigations place the interpretation upon a firmer basis than I had been able to establish in the early reconnaissance, and I accept his statement of the sequence of topographic cycles, as set forth in this paper, as the correct one for the province of the eastern spurs of the Cascade Range and the Columbia basalt plateau. The nonexistence of the Methow Peneplain in that area he proves by detailed observations and strongly supported alternative explanations of the flats, which I thought represented it. They appear to be effects of structural control rather than of peneplaination. We both recognize the Entiat stage as a matureland incised in the supposed Methow. We agree as to the features which represent it west of the Columbia, but neither could identify it on the basalt plateaus. Here I took the wrong road. Thinking that the broad flats of the plateau were of Methow age, I had to find Entiat slopes below them or to explain why they were not there. Unable to discover them, I tried to explain their absence, but I was never convinced by my own suggestions. When Waters traced Entiat surfaces under the basalt flows he proved that my mistake began where I failed to recognize structural control in lieu of Methow plaination.

Thus the Entiat topographic cycle shifts from a supposed Pliocene or younger date to a pre-basalt Miocene position and is the oldest cycle identified in the area studied by Waters. Is it then the oldest identifiable in the heights of the Cascade Range, or is there an older erosion surface to be seen there?

Looking back over a vista of 40 years, I appreciate the unlike developments of diverse regions as I did not in the late 1890's and would make distinctions I did not then recognize. During the field seasons of 1894-1895, and 1896 I observed the summits of the Cascade Range from many points of view and was impressed by the approximation to uniform altitudes among them. Russell saw the same uniformity. Both of us were influenced by the concept of the peneplain so cogently reasoned by Davis upon the evidence of level crests of the Appalachians. Neither of us then realized the difference of orogenic histories and physiographic cycles in Atlantic versus Pacific regions; the former long dormant, possibly entering upon renewed activity, the latter intensely active since Cretaceous time and showing no signs of decadence. Now that I have seen North Atlantic lands from Florida to France and Pacific lands from Patagonia around to New Zealand, I recognize that peneplains are typically Atlantic, maturelands typically Pacific.

In the light of this general contrast, I now entertain the concept of a Methow matureland, in lieu of the supposed Methow Peneplain, as that old topographic surface which we see in the Cascade summits. I would grant a certain degree of factual background to Dawson, Daly, and others, who have argued that an approach to uniformity of elevations may be attained among summits above sea level or grade plain, provided the topography has repeatedly been reduced to a mature aspect, presenting wide valleys and residuals, though it may not at any stage have been peneplained. The great extent of smooth profiles and the inequality of conspicuous peaks, standing out as monadnocks or inselbergs (Willis, 1936), both of which are common in Pacific coast ranges, may best be explained by thus recognizing the effect of brevity of quiet intervals between repeated episodes of activity, during the post-Jurassic orogenic pulses of that vast basin.

If now we substitute the concept of a Methow matureland for that of the Methow Peneplain, do we identify the Methow stage in the high Cascades with the Entiat stage of the eastern slope? Possibly, but not necessarily.

Let us go back to the Miocene landscape, before the Cascade Range began to rise to the heights it has since attained. It presented a fairly broad plateau, surmounted by many residuals. If that was the Methow matureland, a predecessor of the Entiat, then there ensued an initial upwarp, in which Entiat valleys were sculptured. Growing retrogressively headward they may or may not have succeeded in removing the older plateau on the divides. If they did not everywhere do so, and if later canyons have not pushed back to the crest, then a remnant of Methow age may survive in a pass here and there. What is the flat meadow of Snoqualmie Pass, for instance? Is it a glacial deposit or an Entiat valley floor, or a Methow remnant, or something not yet thought of?

These speculations point toward the inference that the accordance of summit levels in the high Cascades is not a result of one-time peneplaination of the perfected Atlantic type but that it may be regarded as suggesting two conditions of mature topography, namely (1) An approach to equality of heights among some residuals during successive cycles of erosion, and (2) survival of plateau surfaces or wide valley floors at headwaters. The latter may represent Methow or Entiat, or here one and there the other. The distinction will be difficult to establish. If, perchance, all high-level flats and smooth profiles are traced to recognized Entiat surfaces, then the Methow stage recedes into the pre-Entiat past, but even so the name may well be retained for that earlier, vanished landscape.

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